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LEACHING OF WISCONSIN GLACIAL GRAVELS IN EASTERN NORTH AMERICA

By Paul MacClintock

Abstract

The depth to which a drift sheet has been leached of its limestone depends on many factors, one of which is length of time. If other variables could be reduced, depth of leaching might be used as criterion of age. Porous gravel in flat-topped hills was measured, and carbonate content ascertained. Observations fall into a pattern of three belts: a southern belt leached 11–13 feet, a middle one leached 5–8 feet, and a northern belt leached 1–4 feet. Within each belt where the drift contains limestone, the depth of leaching is fairly constant even though percentage of carbonates varies considerably. The study therefore suggests that the depth of leaching of gravel is a fairly good criterion of age, and that the southern belt is Tazewell, the middle belt is Cary, and the northern belt is Mankato.

This would make the Olean of New York and the drifts of northeastern New Jersey and Long Island Tazewell, the Valley Heads, the Binghamton, and the "upper" drift of New England and coastal Gaspe Cary, and the drift south of Lake Ontario and north of the Adirondacks and southern Quebec Mankato.

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Introduction

Four chronological subdivisions of the Wisconsin glacial stage are in use in the Upper Mississippi Valley region:

4. Mankato substage
   Two Creeks interstadial
3. Cary substage
   Brady interstadial
2. Tazewell substage
   First interstadial
1. Iowan substage

These are based on stratigraphic as well as topographic evidence (Leighton, 1933; Frye and Leonard, 1949; Condra *et al.*, 1950; Schultz *et al.*, 1951; Richmond, 1950). In places moraines may be continuously traced and found to overlap so that chronology may be determined. Identifiable loess sheets occur between tills, and thin soils and peat deposits have been recognized between drift sheets. But these criteria of subdivision have not as yet been found in the northeastern part of the

continent. Furthermore, instead of continuous sheets of drift spread across flat to gently rolling country, this eastern area contains only patchy masses of drift on rough to mountainous terrain. The drift is thin or absent over hilltops and thick but isolated in valley bottoms. Exposures are few and far between in the large territory of the Wisconsin Glaciation. Since streams have not had time to cut valleys, manmade exposures along highways, railroads, gravel and barrow pits offer the only exposures for the Pleistocene geologist to study. Furthermore, the stagnation of the ice sheets in the rough country has left patches of kame terrace and eskerine material rather than continuous marginal moraines with aprons of outwash sloping away.

The outermost border of Wisconsin drift forms an easily recognized moraine, "The Terminal Moraine" of Lewis (1884) and Salisbury (1902). North of this margin, however, the borders of the Cary and the Mankato drift sheets are very faint and highly irregular because the ice, at its maximum extent in each substage, stagnated in the rough topography and parted into irregular ice masses. The interstadial intervals were very short compared with the major interglacial intervals of the Pleistocene, so that only thin and "infantile" soils were formed, and these only in the most advantageous places and on parent material most easily converted to a soil profile (Schultz, 1951, p. 7; Richmond, 1950, p. 1497). These conditions have not been found in New England. Complications of overriding and irregular burial and/or discontinuity, as described by White (1952, p. 1312) have confused the issue.

The current study is an attempt to find a test sufficiently delicate to subdivide the Wisconsin drift sheet into its substages and yet applicable over the great variety of topographic, lithologic, and climatic conditions. It is proposed that the depth to which the carbonates have been leached from calcareous gravel offers such a test and, within reasonable limits, is a measure of relative age of drift sheets. This idea may furnish a guide and incentive to future detailed stratigraphic work.

Acknowledgments

Field expenses have been derived from three sources:—(1) the generosity of Childs Frick; (2) funds of the Higgins Trust; and (3) Department of Geology, Princeton University. Field assistants have been L. R. Peltier, Albert Pittis, Copeland MacClintock, and John Ostrom. Field discussions have been held with George White, Earl Apfel, Kemble Widmer, Charles Denny, and Walter Lyford. Innumerable local people have been helpful, and particular thanks go to Dr. Stanley Walker of Presque Isle. Two of the pebble counts in Vermont were furnished by J. C. Stone (1953, Glacial geology of Hardwick quadrangle, Vermont: Senior thesis, Princeton University).

Depth of Leaching

The depth of leaching of the carbonates has been used for many years as a criterion of age of drift sheets. Flint (1949), however, warns against accepting depth of leaching as a simple test of age because of the many variables involved. He concludes that it is a significant test for separating major stages of the Pleistocene but is not reliable for separating the substages of the Wisconsin. This discussion considered only the leaching of till.

Many conditions of course influence the depth of leaching. Some of these are interdependent, whereas others seem to be straight-line functions.

(1) Time. With the passage of time leaching progresses deeper and deeper.

(2) Size of the particles of carbonate. The larger the piece of limestone the longer it takes to dissolve it away. However, the converse, *i.e.* that the finer the material the faster the leaching, encounters the complication of fine material being less permeable. There seems to be an optimum size-permeability ratio which gives most rapid leaching. Field observation shows that porous silty till or loess is leached most rapidly if all other conditions remain constant.

(3) Rate of percolation. Size of the particles, topographic position, character of rainfall, whether gentle or torrential, and the type of vegetal cover influence infiltration and percolation.

(4) Amount of rainfall. Undoubtedly this is a factor in depth of leaching.

(5) Temperature. One might at first think that leaching would be faster in southern and warmer latitudes, but, since CO_2 is more soluble in cool water, the relation of temperature zones to the leaching of carbonates is far from simple. Furthermore, the type of vegetation may influence rate of leaching more than the temperature alone.

(6) Biochemical conditions. Organic acids speed up leaching.

(7) Soil. Type of soil and special soil conditions such as impervious pan horizons, are of major significance.

(8) Topographic situation. Uplands leach faster than lowlands at or near the water table. Slopes and steep hilltops are eroded as leaching takes place so that leaching is shallower.

(9) Amount of carbonate material. Other things being equal, the greater the amount of carbonate originally present the shallower the leaching.

Method of Field Observation

General Statement

In order to eliminate the variables insofar as possible within the limitations of field work, and to produce comparable data for study, observations were made as follows:

Porous gravel in active gravel pits was tested for depth of leaching below a horizontal, soil-covered surface well above the water table. In this way were eliminated the varying rates of percolation occurring in till as well as the possible erosion of leached material from the surface. It is necessary to study an active gravel pit if depth of leaching is more than just a few feet because, within a day or two after the power shovel stops digging, the leached gravel slumps over the unleached gravel and makes a valid test of leaching impossible. Consequently, in the current study, exposures where leaching is only 1–2 feet deep are easily found; exposures of 6 or 7 feet of leaching are abundant but often require a little digging with a hand shovel; but exposures where leaching has progressed to 10 or more feet require the active co-operation of a power shovel and, hence, are much less numerous. To standardize the size of particles, the sandy matrix of the porous gravel was tested, and the depth of leaching placed where the sand effervesces to dilute HCl or "sizzles" when held to the ear. In most gravel pits, depth of leaching varies somewhat from place to place within the pit. At one spot it might be 5 feet, but 30 feet away it might be 7 feet. After making tests at several places in a pit the shallowest and the deepest leaching was recorded. The amount of carbonate material originally present in the gravel was ascertained by making pebble counts of the unweathered calcareous material. It is assumed tentatively that rainfall has been roughly uniform within the region and that well-drained soils on gravel are uniform enough in soil type for present purposes. The question of the variation in temperature from southern New England to Canada and its possible influence on rate of leaching lies beyond this present discussion. Field observations indicate no gradual transition in depth of leaching from south to north as might be expected if temperature gradient were of major importance.

Secondary Lime

In most gravel pits secondary lime has been deposited on stones. In places, enough of this lime is deposited to cement the pebbles into masses of conglomerate 8–10 feet across. In many places, individual pebbles are crusted with the white $CaCO_3$ precipitate. Commonly, limestone pebbles acquire more encrustation than do the adjacent siltstones or sandstones, but probably the most characteristic occurrence of this white lime deposit is on the under side of pebbles. The top of the pebble may be free of lime deposit even though the bottom may have enough to cement several smaller grains to it. Two hypotheses might explain these pendant lime crusts: (1) They are the result of precipitation exclusively under the stone; some type of stagnant condition forms under the stone which would cause $CaCO_3$ precipitation. (2) $CaCO_3$ is precipitated on the top as well as side and bottom of the stone and later this crust is leached off the top of the pebble, but not leached out from under it; the pebble acts as a sort of roof to protect material im-

mediately below it. Under (1) the depth of these lime crusts below the surface of the ground is a problem of ground-water hydrology and physical chemistry. Under (2) it is a measure of the depth of leaching.

However the size of the stone is significant. At the State-line gravel pit 1 mile north of Canaan, Connecticut, the gravel is leached to a depth of 5 feet but, adhering to the under side of a flat stone 8 by 6 by 2 inches, a deposit of $CaCO_3$ is only 3.5 feet from the ground surface. In the Lee Brothers' pit, 4 miles north of East Stroudsburg, Pennsylvania, lime is present 8 feet below the surface under a 4 by 5 foot boulder, whereas the pit as a whole is leached 25 feet. If the lime crusts are to be used in measuring the depth of leaching, only those crusts found on small stones of an inch or less could be used with any degree of confidence. It is interesting that these lime ($CaCO_3$) crusts form from the leaching of both limestone and dolomite pebbles. This fact has been of great assistance in field work because the presence of carbonates in a gravel deposit can be inferred at once by finding crusts on the floor of the pit, even when dolomitic siltstones may dominate the content of calcareous material in the gravel. On the other hand the common presence of secondary lime on pebbles of the gravel has made it necessary to crack open every pebble in making pebble analyses and to test the freshly broken surface under a magnifying glass to be sure that effervescence does not come from secondary lime on the outside of the pebble.

Many glacial geologists and pedologists have been consulted as to whether or not the lime crusts could be used in testing depth of leaching. No strong objection has been voiced, and Guy D. Smith expressed the opinion that they would probably serve to limit the depth of leaching (Personal communication, 1949).

By careful selection of exposures in the field, an attempt is made to standardize porosity, topography, and ground-water conditions, but there still remains the question as to whether original content of carbonates determines depth of leaching. When the depth of leaching is plotted against the percentage of carbonate in the original gravel it is seen that, whereas carbonate content does play a role, it is not the controlling factor. Drift may be leached 1½ feet, even though carbonate content ranges from 11 per cent to 70 per cent, and leached 6 feet though carbonate ranges from 9 per cent to 62 per cent and leached 12 feet though carbonate content ranges from 5 per cent to 35 per cent.

The graph shows three groupings, an upper one with its node at about 1½ feet of leaching, a middle one with a node at about 6 feet, and a lower one with a node at about 12 feet. Since the gravel deposits where these measurements were made are in well-defined geographic areas without overlap or contamination, it is proposed that they represent different drift sheets of different substages of the Wisconsin. It is further proposed that the drift leached 10–15 feet is the Tazewell, that leached 4–8 feet the Cary, and that leached 1–3 feet the Mankato.

Special Cases

Two special cases have been encountered in the field where depth of leaching is less than normal: (1) At the top of steepish kames or eskerine hills erosion, mass movement, and frost action have removed weathered material so that carbonates are found abnormally near the surface. This is well shown in the kames 2 miles north of Vernon in northern New Jersey. Calcareous material is found within a foot of the surface, and limestone pebbles may occasionally be seen in the grass, whereas the normal leaching of this region is 5–7 feet. (2) In places gravel contains a large percentage of carbonate (50–60 per cent or more). Such places are the Airport pit, Elmira, New York (leached 3½ feet, carbonate 54 per cent); the Great Meadows, New Jersey, kame area (leached 3–4 feet, carbonate 69 per cent); and North Church, New Jersey (leached 3½ feet, carbonate 60 per cent). Surrounding exposures record 5–7 feet of leaching.

Tedrow (Personal communication, 1953) suggests that a large ratio of dolomite to limestone might retard leaching.

Generalized Map of Occurrences

The leaching data plotted on a map falls into three belts or zones—a southern belt where

leaching is 11–13 feet, a middle belt with leaching about 5–8 feet, and a northern one leached 1–4 feet (Fig. 2). These would then be the Tazewell, the Cary, and the Mankato drift sheets respectively. This map may best be described in sections.

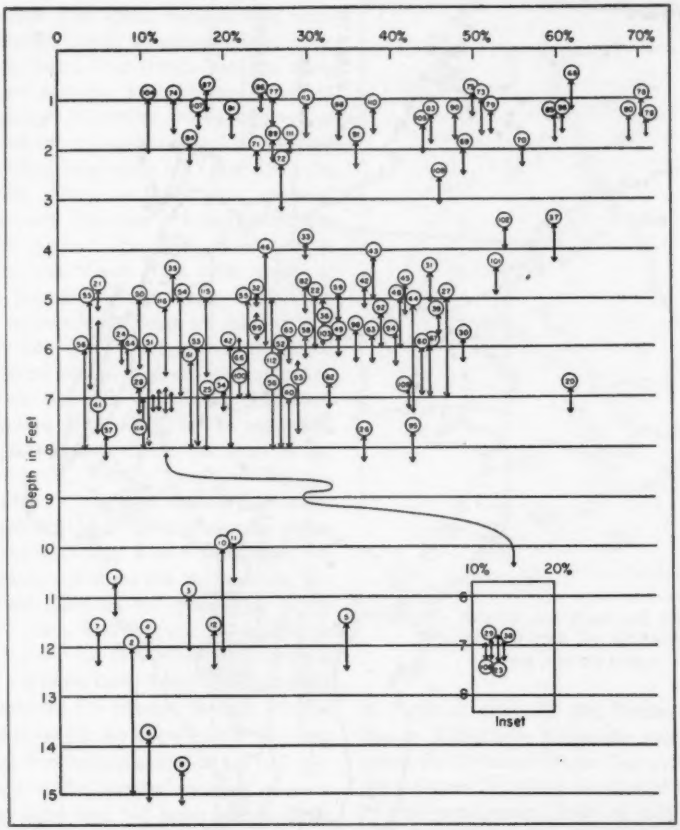

FIGURE 1.—GRAPH OF DEPTH OF LEACHING PLOTTED AGAINST PER CENT OF CALCAREOUS MATERIAL
Vertical lines show shallowest to deepest leaching found in each gravel pit.

DELAWARE VALLEY

The current study of the leaching of glacial gravel started in the attempt to map the terraces of the Delaware Valley. The 20-foot terrace is in many places leached 6–8 feet, and the 50-foot terrace is leached 10–15 feet in at least three localities. These observations led to detailed study of the Delaware Water Gap area where the Wisconsin moraine crosses the river at Belvidere, New Jersey. The gravel of the moraine west of Bangor, Pennsylvania (Fig. 3), contained no carbonates to a depth of at least 10 feet, whereas the moraine just east of it, through Portland, Pennsylvania, is calcareous below 5 or 6 feet. Except for carbonates, the lithology of the two moraines is the same. Seemingly these moraines are the product of two episodes of ice movement over the same terrain but at different substages of the Wisconsin: the Bangor moraine is older, and leached 10–15 feet, whereas the younger Portland moraine is leached only 5–6 feet. One mile southwest of Belvidere, New Jersey, the Port-

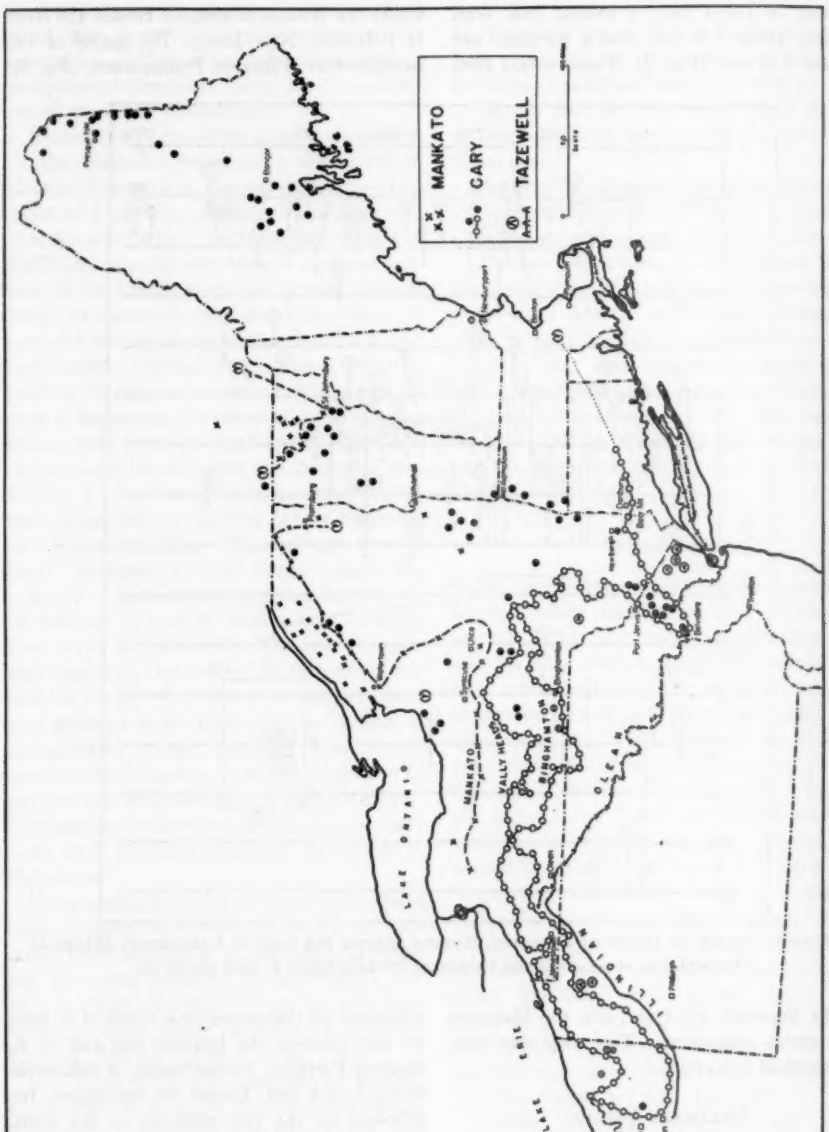

FIGURE 2.—MAP OF LEACHING TESTS AND PROPOSED WISCONSIN CORRELATION
Plotted on and traced from U.S. Geological Survey 1:1,000,000 map base.

land moraine overrides and buries the Bangor moraine and becomes the outer margin of the Wisconsin as far as Netcong, about 25 miles to the northeast, where the more deeply leached drift emerges and again becomes the outer margin through Dover, Morristown, Metuchen, and Staten Island. Four critical localities along the Bangor moraine may be mentioned. A poor exposure 3½ miles west-southwest of Belvidere shows calcareous gravel thrown out of a groundhog hole which is 15 feet down the valley side. Since no calcareous material anywhere along the slope is found above this level it is inferred that the gravel here is leached to this depth. One mile south of Nazareth Junction a fresh slump in the railroad cut exposes fairly dense till leached to a depth of 9 feet. Since observations throughout eastern United States have shown that gravel leaches faster than till, it is inferred that this 9 feet of leaching of till would be the equivalent of 11–14 feet in gravel. On the crest of the Bangor moraine 3 miles north of Bangor, Pennsylvania, a roadside exposure contains deeply weathered and etched calcareous stones about 6 inches in size, 8 or 9 feet below the ground surface. The matrix is, however, entirely leached. Just off the quadrangle to the northwest, the Kellersville, Pennsylvania, pit gives a good test of 11–15 feet of leaching in gravel. Likewise three type localities might be mentioned for the younger moraine. Pits in the kames along Jacoby Creek half a mile west of Portland, Pennsylvania, are leached to 5 feet. At the crest of the Portland moraine where it crosses the older one, 1¼ miles southwest of Belvidere, a pit gives a good test of 5½ feet of leaching. The big pit, likewise on the crest of the moraine, 2 miles east of Belvidere, is also leached 5 feet.

It became evident that the two depths of leaching found in the Delaware Valley terraces had their inception in the two drifts at Belvidere (Fig. 4), and hence the terraces may be dated within the Wisconsin chronology.

In order to correlate these local episodes to the standard Wisconsin section, observations were spread over northern New Jersey and through New York State to Ohio.

FIGURE 2.—MAP OF LEACHING TESTS AND PROPOSED WISCONSIN CORRELATION
Plotted on and traced from U.S. Geological Survey 1:1,000,000 map base.

NORTHERN NEW JERSEY

The two depths of leaching are found northward into New York and westward into Pennsylvania. Northeastward from the town

FIGURE 3.—BANGOR AND PORTLAND MORAINES ON DELAWARE WATER GAP QUADRANGLE
Arrows indicate striae.

of Portland (Fig. 5), the Portland moraine crosses Kittatinny Mountain about 7 miles above the Delaware Water Gap and lies along the Delaware Valley at the base of the Pocono Plateau escarpment. Drift of this episode is difficult to follow along this stretch of its boundary because it was heavily contaminated by local material and older Wisconsin drift, washed down from the Pocono highland. The intermingling is seen to a lesser extent where the southeast edge of the ice lobe lay against the somewhat lower "Jersey Highlands" of Sparta, Hamburg, and Wawayanda mountains, and the highlands of Bear Mountain.

The northeastern part of New Jersey contains only the older Wisconsin drift as shown by 11–15 feet of leaching. The moraine of this older episode crosses Staten Island and becomes the Harbor Hill moraine of Long Island.

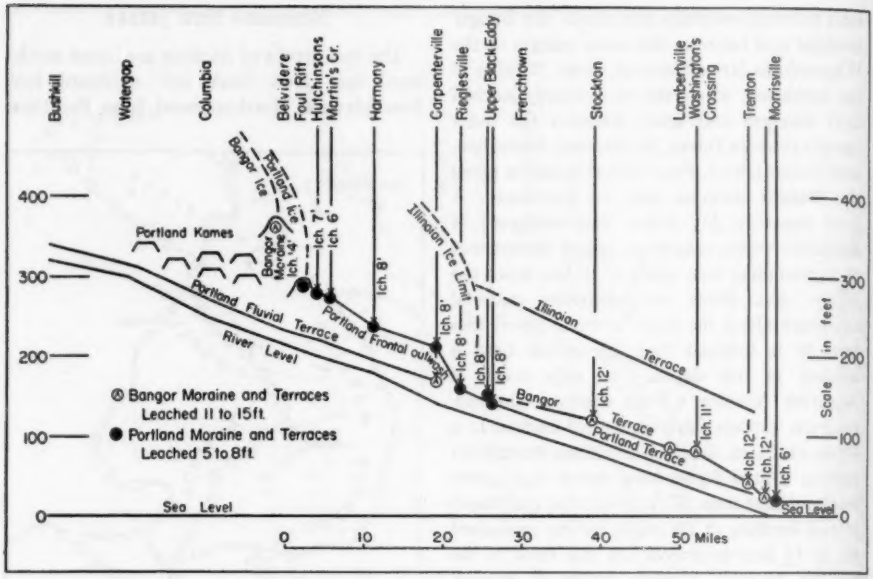

FIGURE 4.—TERRACES OF DELAWARE RIVER

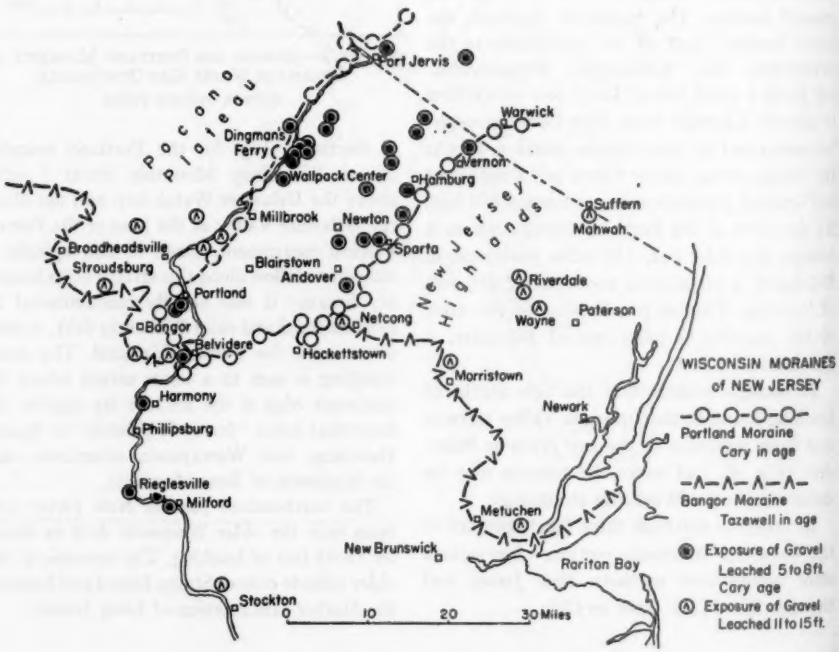

FIGURE 5.—WISCONSIN DRIFT IN NORTHERN NEW JERSEY

When traced northward, the Portland drift becomes the "younger" drift of Rich (1935, p. 130) who found that ice of this episode pressed into the Catskills from the north and east, did not overtop this highland, but left only earlier Wisconsin drift in the western and southwestern part. In the present study this older drift is found to be leached 12–15 feet at Roscoe and Horton, New York, southwest of the Catskills, and hence is correlated with the Bangor moraine of the Delaware Water Gap region. This Bangor drift has been mapped across the Pocono plateau and becomes the Olean drift of MacClintock and Apfel (1944).

Finger Lakes Region

Only reconnaissance observations were made in this important region. However, even this limited work has produced significant facts. The Binghamton drift at Elmira, Owego, Binghamton, Green, and Cobleskill is leached 5–8 feet. The Valley Heads moraine needs much more study, but at Tully, Scott, Malloyville, Odessa, and south of Ithaca, depths of 5 or 6 feet have been found. Between the big Valley Heads moraine at Tully, New York, and the Binghamton moraine at Oneonta, New York, good tests of 5–7 feet of leaching are found at South Edmonds, New Berlin, and South New Berlin along route New York 8. It suggests that the Binghamton and Valley Heads moraines beling to the same substage of the Wisconsin—the early and late Cary respectively (MacClintock and Apfel, 1944, p. 1162). At the north end of the Finger Lakes, however, the evidence is quite different. At the north end of Skaneateles Lake, 1 mile east on route U.S. 20, a pit is leached only 18 inches; and 3 miles west, also on route 20, another pit is also leached, 18 inches. A patch of good terminal moraine topography lies at the east end of Seneca River Bridge, at the northern end of Lake Cayuga; here the till is leached only 18 inches; 2 miles eastward, in the same patch of moraine, a gravel pit is leached only 1 foot. Cayuga Lake is dammed by drift of shallow leaching. Gravel pits were not found at the immediate north end of Seneca Lake, but farther north at West Junius, the Ontario Sand and Gravel Company pit effervesces 12–18 inches below the surface. Half a mile southeast of Canandaigua, New York, the strong moraine which forms the dam at the north end of Canandaigua Lake contains a gravel pit leached 18–20 inches. The morainal topography is flattened where the lake spilled northward in the area now occupied by the city park. It

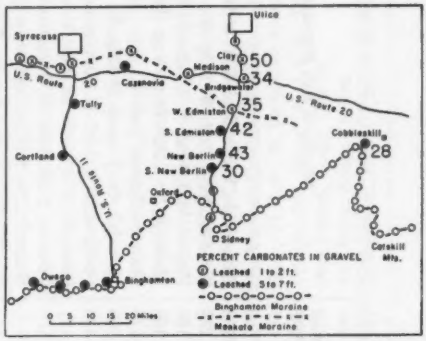

Figure 6.—Detailed Map of Small Area South of Utica Showing Two Depths of Leaching in Gravel of Similar Carbonate Content

appears that all the Finger Lakes have been dammed by Mankato drift.

Westward from Canandaigua, Mankato gravel is encountered at the following places: (1) The eastern outskirts of East Bloomfield on route U.S. 20; (2) Mr. Shattuck's pit on route New York 5, 3½ miles southeast of Caledonia, New York; (3) 2 miles south of Rega on route New York 36; (4) near Ivison's pit near Bergen, New York, route New York 33; and (5) Corfu route New York 33.

Whereas some of these exposures do contain a large percentage of carbonate material, this fact alone cannot account for the leaching which is much shallower than that encountered a few miles to the south (Honeoye, N. Y., leached 7–8 ft.).

Eastward from Skaneateles Lake, Mankato moraine forms a dam across Otisco Lake Valley, through which Ninemile Creek has cut its postglacial valley. A gravel pit on the north flank of this morainal embankment is leached 1½ feet. At Cedarvale, in the northeast part of Skaneateles quadrangle and northwest part of Tully quadrangle, Mankato terminal moraine forms a considerable embankment. Pits southwest and east of Indian Village Tully quadrangle, are leached 1–2 feet. Pits at Jamesville are likewise leached only 1–2 feet.

The eastern Finger Lakes region supplies

TABLE 1.—PEBBLE ANALYSES

Station	Depth of leaching in feet	Limestone	Dolomite	Total carbonates	Sandstone	Sandstone and siltstone	Siltstone	Quartz and/or chert	Quartzite	Igneous	Schists	Igneous and metamorphics	Total stones	Location on State Highways Maps
1	11	..	..	7	..	20	69	.5	2	..	..	..	1168	Kellersville, Penna. 5 mi. S.W. of Stroudsburg (Fig. 5)
2	12–15	..	..	9	..	..	1	.5	..	89	..	..	217	Netcong, N. J. 1 mi. W. (Fig. 5)
3	11–12	..	..	16	..	14	2	..	2	42	..	22	198	Wayne, N. J. 1 mi. N.E. route 23 (Fig. 5)
4	11	..	..	10.7	9	..	2.8	2.6	20	54.6	..	..	150	Mahwah, N. J. Sam Brean's pit (Fig. 5)
5	11.5	16	19	35	..	56	..	1	7	..	..	..	245	Shawnee, Penna. 3 mi. E.
6	14–15	..	..	11	..	7	..	..	16	65	..	..	230	Riverdale, N. J. route 23 (Fig. 5)
7	12	..	..	3	..	33	..	15	33	16	..	..	130	Stockton, N. J. 1 mi. N. 40 ft. terrace (Fig. 4)
8	15	..	..	15	..	..	..	..	..	..	..	..	..	Olean, N. Y. 1 mi. W. on route U.S. 17 (Fig. 2)
10	10–12	..	..	20	..	..	..	..	..	..	..	..	..	Panama, N. Y. 3 mi. W. Jamestown Quad. (Fig. 2)
11	10	..	..	21	..	..	..	..	..	..	..	..	..	Watts Flatts, N. Y. Chautauqua Quad.
12	12	..	..	19	..	..	..	..	..	..	..	..	..	West Valley, N. Y. 4 mi. W., Ellicottville Quad.
20	7	..	..	61.8	11.8	..	4.3	14.7	7.5	..	..	..	161	Hutchensons, N. J. 60 ft. Terrace (Fig. 4)
21	5	..	..	7	..	47	..	25	5	6	10	..	128	Hudson, N. Y. 19 mi. E. on route 23
22	5–6	..	..	31	..	8	..	16	10	16	19	..	100	East Calais, Vt. 1 mi. N. on route 12
23	7	4	9	..	..	..	3	..	66	..	..	18	239	Bristol, Vt. 5 mi. E.
24	6	..	..	8	..	29	..	..	57	..	..	6	231	Morrisville, Penna. 2 mi. S.W. 20 ft. terrace (Fig. 4)
25	6–7	15	3	18	..	44	..	..	36	..	..	1	199	Wallpack Center, N. J. Wallpack Quad. (Fig. 5)
26	8	14	23	57	..	33	..	2	27	1	..	..	180	Upper Blk. Eddy, Penna. 25 ft. Terrace. (Fig. 4)
27	5–7	5	42	47	..	24	..	2	23	..	..	4	220	Martins Creek, Penna. 60 ft. Terrace
28	7	7	4	11	..	73	..	..	16	..	..	..	155	Dingman's Ferry, Penna. (Fig. 5)
29	7	11	2	13	..	50	..	1	36	..	..	1	338	Millbrook, N. J. 1½ mi. S.W. Wallpack Quad. (Fig. 5)
30	6	..	..	49	..	..	24	..	12	15	..	..	100	Vernon, N. J. Essex County (Fig. 5)
31	4.5	12	23	45	14	..	6	..	42	..	1	..	208	Portland, Penna. 1½ mi. W. (Fig. 5)
32	5	..	..	24	..	63	..	8	..	..	..	4	210	Pennellville, N. Y. ½ mi. S.
33	4	30±	..	..	..	..	..	..	..	..	..	..	..	Pennellville, N. Y. 3 mi. N. on route 49
34	7	..	..	20±	..	..	..	..	..	..	..	..	..	Fulton, N. Y. route 57
35	4.5–5	6	8	14	..	79	..	..	4	3	..	..	139	Mexico, N. Y. 5 mi. S. E. route 69
36	5–6	32	10	42	..	35	..	..	16	..	..	4	224	Rome, N. Y. route 69

36	5–6	32	10	42	..	35	..	..	16	..	..	4	224	Rome, N. Y. route 69
37	4	..	..	60	..	13	..	3	10	..	..	14	156	North Church, N. J. route 31
38	7	4	9	13	..	..	3	..	66	..	..	18	249	Bristol, Vt. 2 mi. E. route 17
39	5.5	..	..	45.8	6.6	..	1.1	10.2	33.9	2.6	..	..	274	Rieglesville, Penna. 25 ft. terrace (Fig. 4)
40	6–7	27	1	28	4	..	65	1	1	1	..	..	170	West Richmondville, N. Y. route 7 (R.R. cut)
41	5–8	..	..	5	..	..	..	..	61	19	15	..	171	Canaan, Conn. 1 mi. N.
42	5	..	..	36	..	..	..	..	42	13	9	..	136	Williamstown, Mass. 4 mi. N. route 7 (Fig. 2)
43	4–5	5	33	38	7	..	13	28	10	..	4	..	131	Portland, Penna. 1 mi. W. (Fig. 5)
44	5–7	37	6	43	..	37	..	7	11	..	..	1	375	New Berlin, N. Y. 3 mi. S. route 8 (Fig. 6)
45	5	37	5	42	..	55	..	3	..	..	..	..	311	South Edmeston, N. Y., route 8 (Fig. 6)
46	4–6	18	7	25	7	..	31	..	17	..	..	20	213	Saratoga Springs, N. Y. 5 mi. S. route 9 (Fig. 2)
47	6–8	19	2	21	1	..	6	1	18	..	..	53	325	Glens Falls, N. Y. 6 mi. S. (Fig. 2)
48	5–6	..	..	42	..	54	..	..	4	..	..	..	123	Houlton, Maine. W. on route 2
49	6	..	..	34	..	62	..	..	3	..	..	..	120	Monticello, Maine. 10 miles W.
50	5–7	..	..	10	..	..	60	15	10	4	..	..	165	Newburg, Maine. S.W. Bangor route U. S. 202
51	6–8	..	..	10.7	..	..	50	10	22.5	5	..	..	111	Plymouth, Maine. W. of Bangor route 7
52	6–8	27	2	29	..	..	..	9	34	6	22	..	158	Troy, Maine. S.W. of Bangor route U. S. 202
53	6–8	..	..	17	..	..	36	5	30	11	..	..	138	St. Albens, Maine. W. of Bangor route 43
54	5–7	..	..	15	..	..	64.6	2.2	13.5	4.5	..	..	133	Garland, Maine. 25 mi. N.W. Bangor
55	5–7	22.8	.7	23.5	..	..	..	1.4	32.1	4.2	38.6	..	140	Corinth, Maine. N.W. Bangor route 43
55a	5–7	..	..	4.1	..	..	51.1	14	12.2	18.3	..	..	213	Sherman Mills. Maine 1 mi. N. route 11
56	6–8	..	..	2.3	..	54.3	..	50	18.3	16.8	..	..	184	Lincoln, Maine. route U. S. 2
56a	6–8	..	..	27	..	..	57	12	..	7	..	..	146	Island Falls, Maine. E. on route U. S. 2
57	8	..	..	6	..	80	..	..	9	4	..	..	138	Houlton, Maine. 9 mi. W. route U. S. 2
58	6	..	..	30	..	..	60	10	..	..	..	..	153	Mars Hill, Maine. 5½ mi. N.E. (Clyde York pit)
59	5–6	..	..	34	..	..	55	..	8	3	..	..	145	Easton Corner, Maine. 2 mi. E. route U. S. 1A
60	6–7	..	..	44	..	..	48	..	8	..	..	..	168	Carabou, Maine. 1 mi. E. (Gould pit) route U. S. 1
61	6–8	..	..	17	..	70	..	..	13	..	..	..	157	Hamlin, Maine. 1 mi. E. route 165
62	7	..	..	33	..	13	..	27	13	8	5	..	157	Cap Chat, Gaspe. 1 mi. W.
63	6	..	..	38	..	..	17	2	26	12	5	..	129	St. Armond, Quebec. 10 mi. E.
64	6	..	..	8	..	..	8	14	21	15	34	..	111	Stow, Vt. 2 mi. N. Moss Glen route 100
65	6	..	..	28	..	61	..	..	7	4	..	..	131	Presque Isle, Maine. 3 mi. S.W. Echo Lake
66	6	..	..	22	..	..	18	..	8	11	33	..	136	Port Daniel, Gaspe, Quebec
67	6–7	..	..	45	..	37	..	1	1	8	5	..	100	Fox River, Gaspe, Quebec
68	1–1.5	..	..	62	..	32	..	..	4	2	..	..	129	Bridgewater, Maine. 3 mi. N.E. route U. S. 1

TABLE 1 (*Continued*)

Station	Depth of leaching in feet	Limestone	Dolomite	Total carbonates	Sandstone	Sandstone and siltstone	Siltstone	Quartz and/or chert	Quartzite	Igneous	Schists	Igneous and metamorphics	Total stones	Location on State Highway Maps
69	2–2.5	..	..	49	..	41	..	..	10	1	..	..	176	Westfield, Maine. Aroostook Co. route U. S. 1
70	2.25	16	40	56	4	..	1	1	18	..	..	20	270	Saratoga, N. Y. 5½ mi. W. route 29
71	2.5	..	..	24	..	..	..	..	12	64	..	..	125	Danbury, Conn. 6 mi. N. Elwood pit route U. S. 7
72	2.5–3	..	..	27	..	21	47	15	6	23	8	..	129	Windsor Quebec, route 5
73	1–1.5	..	..	51	..	..	..	..	..	..	..	..	136	Fort Fairfield, Maine. route U. S. 1A
74	1–1.5	..	..	14	79	..	..	..	..	8	..	..	114	Plattsburg, N. Y. 5 mi. No. Ingreham Esker
75	1±	7	43	50	20	..	20	1	..	..	..	8	366	Bergen, N. Y. 5 mi. W. route 33
76	1±	62	35	97	..	2	..	1	..	..	..	..	88	Manlius, N. Y. route 92
77	1.5	23	3	26	..	36	..	3	32	..	..	2	239	Bridgewater, N. Y. 2 mi. N. route 8
78	1.5	..	..	88	5	..	..	..	..	6	..	..	111	Norfolk, N. Y. 2 mi. S. route 56
79	1.5	..	..	51	15	..	15	..	13	10	..	..	103	Ogdensburg, N. Y. 3 mi. S.E.
80	1.5	..	..	69	12	..	..	..	8	10	..	..	106	Ogdensburg, N. Y. 4 mi. S.E. route 68
81	1.5	..	..	21	40	..	19	..	6	15	..	..	248	Waterman Hill, 1 mi. E. Canton Quad.
82	5	..	..	30	..	..	..	1	43	..	..	22	177	West Pierpont, N. Y. Canton Quad.
83	1.5–2	..	..	45	..	..	4	..	38	12	..	..	184	Potsdam, N. Y. 3 mi. S. Potsdam Quad.
84	2	..	..	16	..	..	..	..	55	28	..	..	135	Colton, N. Y. 1 mi. S. Potsdam Quad.
85	1.5	..	..	60	..	..	6	..	27	6	..	..	125	Madrid, N. Y. 2 mi. N. Canton Quad.
86	1	..	..	25	56	..	14	..	..	6	..	..	218	Malone, N. Y. 2 mi. S. Malone Quad.
87	1±	..	..	18	33	..	6	..	..	43	..	..	139	Colton, N. Y. 2 mi. S.W. Potsdam Quad.
88	1.5	..	..	34	15	..	4	..	..	..	..	48	217	Richfield, N. Y. 1.5 mi. S. route U. S. 11
89	2	..	..	26	49	..	10	..	..	15	..	..	194	DeKalb Junction, N. Y. 4 mi. W. route U. S. 11
90	1.5	..	..	48	25	..	6	..	..	21	..	..	153	Sprague ville, N. Y. 2 mi. W. Gouverneur Quad.
91	2	..	..	36	21	..	7	..	..	36	..	..	151	West Fowler, N. Y. Gouverneur Quad.
92	5.5	17	22	39	7	..	9	..	..	..	..	45	189	Harrisville, N. Y. 3 mi. N. route 31
93	6–7.5	..	..	29	60	..	4	..	..	..	..	8	133	Brainardsville, N. Y. 1 mi. S.E. route 374
94	6	..	..	41	27	..	..	..	..	..	..	33	197	South Russell, N. Y. 1 mi. N.W. Canton Quad.
95	8	..	..	43	17	..	18	..	10	..	..	13	119	Newburg, N. Y. 1 mi. S. route 9W
96	1.5	..	..	60	..	..	14	..	2	23	..	..	166	St. Johnsbury, Vt. 1 mi. N. (Fig. 2)

96	1.5	..	..	60	..	..	14	..	2	23	..	..	166	St. Johnsbury, Vt. 1 mi. N. (Fig. 2)
97	2.5–3	..	..	46	..	..	8	..	11	34	..	..	96	Barton, Vt. 2 mi. E. route U. S. 5
98	6	..	..	36	31	..	13	..	..	20	..	..	187	Hardwick, Vt. 1 mi. W. route 15
99	5.5	..	..	24	..	..	35	..	27	..	..	15	167	East Berkshire, Vt. 5 mi. W. route 105
100	6–7	..	..	22	..	17	..	..	30	32	..	..	145	Hardwick, Vt. 3 mi. W. route 15
101	4.5	..	..	53	16	..	21	..	2	7	..	..	110	Binghamton, N. Y. 5 mi. W. route 17
102	3.5	..	..	54	11	..	33	..	2	1	..	..	138	Elmira, N. Y. 3 mi. N. airport pit
103	5	..	..	32	16	..	39	..	4	4	..	..	133	Elmira, N. Y. 2 mi. S.E. route 17
104	1–2	..	..	11	..	..	..	..	33	27	29	..	100	Lowell, Vt. off route 100
105	1.5	..	..	44	..	..	..	..	18	21	17	..	100	Coventry, Vt. 2 mi. S.W. route U. S. 5
106	7	..	..	11	..	..	..	..	16	42	31	..	100	Barton, Vt. 6 mi. S. route 12
107	1.5	..	..	17	..	..	..	..	30	21	32	..	100	Orleans, Vt. 2 mi. S. route U. S. 5
108	3	..	..	46	..	..	9	12	..	..	33	..	100	Barton, Vt. 2 mi. E. (revisited for check)
109	7	..	..	44	..	..	..	..	16	16	29	..	105	S. Walden, Vt. route 15
110	1.5	..	..	38	..	..	..	16	..	32	14	..	100	Coventry, Vt. 1 mi. N. route U. S. 5
111	2	..	..	28	..	..	..	14	31	5	22	..	100	Johnson, Vt. 1 mi. E. route 15
112	5–6	..	..	26	..	..	..	8	26	15	25	..	100	Jeffersonville, Vt. 2 mi. N. route 15
113	1.5	..	..	30	..	..	..	8	..	25	37	..	100	Sheldon, Vt. route 105
114	7–8	7	4	11	..	73	..	..	16	..	..	..	155	Dingmans Ferry, Penna. (Fig. 5)
115	5–8	15	3	18	..	44	..	..	36	..	..	1	199	Walpack Center, N. J. (Fig. 5)
116	5–8	11	2	13	..	36	..	1	..	..	..	1	338	Millbrook, N. J. (Fig. 5)
117	1.5	..	..	19	57	..	1	..	..	..	..	23	184	Patten's Mills, N. Y. Near Glens Falls route U. S. 9

evidence that percentage of carbonates in the drift does not control depth of leaching. In the region south of Utica, along route New York 8, (Fig. 6) percentage of carbonates ranges from 30 to 50, but the higher percentage does not necessarily accompany shallow leaching.

The Syracuse area presents special problems and needs more detailed study. There is shallow leaching east, west and south of the city but deeper leaching of 5–7 feet is found at several points to the north and northwest. The North Syracuse pit is leached 7 feet. At Pennelville, gravel with 24 per cent carbonate is leached 5 feet. At Fulton, on route New York 57, 7 feet of leaching is found in gravel of 20 per cent lime. Near Mexico on route New York 68, 4.5–5 feet of leaching is found in gravel of 14 per cent lime. At Rome the gravel contains 42 per cent carbonates and is leached 5–6 feet. These observations of deeper leaching lie north of the suggested Mankato border. However, at Callamer, New York, 6 miles northeast of Syracuse and along route New York 515, there is evidence that partly weathered gravel has been overridden and incorporated into younger drift. The upper 4 to 5 feet is disturbed and churned up and, although still calcareous, is studded with partly weathered and etched stones. Likewise, at Marietta, south of Syracuse, partly weathered gravel lies below calcareous lake silts. It is a tenable hypothesis that Mankato ice overrode partly weathered drift north of Syracuse. Such an advance of the Mankato ice would illuminate Fairchild's statement (1909, p. 57):

> "Subsequent to the extinction of Lake Vanuxem, the length of time unknown, the ice front readvanced in the Syracuse district with a consequent redamming of the central New York waters. The result was the re-creation of Lake Vanuxem in rising levels and the union with Lake Warren."

Likewise overriding is suggested in the Champlain valley area because at the northern end of Lake Champlain, near St. Armond, Quebec, leaching of 6 feet is found, whereas at Patton's Mills at the southern end only 1.5 feet of leaching is encountered; this suggests that Mankato ice projected down the Champlain valley but failed to destroy the weathered drift at the north end. White, (1952, p. 1312) has found similar cases in Ohio where overriding has taken place over considerable areas with nondeposition of drift.

Maine and Quebec

Leaching tests were made on gravel deposits in western Massachusetts, in Vermont, and in Maine, wherever calcareous material could be found. As seen on the map (Fig. 2) and in the table (Table 1), the data are very spotty, but they can be considered suggestive if not conclusive. Gravel is leached 5–8 feet within this large territory, covering a large range of lithologies. This fact makes correlation with the mid-Wisconsin Cary substage very attractive. The exceptionally shallow leaching of 2–3 feet in several exposures near Mars Hill, Aroostook County, Maine, is attributed to the very high content of carbonates. Aroonstook county is underlain by limestone.

Two localities in Gaspe display Cary drift: (1) At Port Daniel, midway along the south coast, kame gravel 100 or more feet above sea level and composed of 22 per cent carbonates is leached 6 feet; and (2) at Riviere aux Renards, on the northeast coast, gravel containing 45 per cent lime is likewise leached 6–7 feet.

At Danville, Quebec, along route 5, 4–5 feet of leaching was measured and 3 feet measured at Windsor. However, as previously noted, 6 feet of leaching occurs at St. Armond near the north end of Lake Champlain.

St. Lawrence Lowland

Fifteen observations between the Adirondack upland and the St. Lawrence River show gravel leached 1–2 feet; and five localities in the northern Adirondacks show 5–7 feet of leaching (MacClintock, 1953). This indicates that Cary ice overrode the Adirondacks but that Mankato ice failed to "make the grade" and stood briefly at a position just south of Gouverneur, Canton, Potsdam, and Malone.

Summary

The study of leaching of glacial gravel suggests the following correlations: The Tazewell of Ohio correlates *via* Spring Valley, Pennsylvania (2 miles south on route 8), with leaching of 15 feet; with the Olean of New York; with the Bangor moraines of the Delaware Water Gap area; with northeastern New Jersey; and with the Harbor Hill moraine of Long Island. A natural extension would include the Watch Hill and Southern Cape Cod

moraines and the "Lower" till of Massachusetts. The Cary of Ohio (7–8 feet of leaching at Kinsman, Ohio; at Mesopotamia, Ohio; and at St. Joseph, Ohio) would correlate with Binghamton and Valley Heads drifts of New York (MacClintock and Apfel, 1944, p. 1162), the younger drift of Rich (1934) in the Catskills, the Portland moraine of the Delaware Water Gap, and the "Upper" drift of New England, south to an undetermined boundary near Danbury, Connecticut, Hartford, and Boston (MacClintock and Peltier 1949, p. 1971).

The Mankato margin lay at the north end of the Finger Lakes and along the north flank of the Adirondack upland. It stood across Quebec and extended into Vermont as far south as St. Johnsbury, but fell short of northern Maine and the Gaspe, except possibly for local ice caps on the highlands.

Two cases of pollen analysis are compatible. In the pollen profile of a peat bog in High Point State Park, New Jersey, Niering (1953, p. 141) finds cold assemblage at the base overlain by warmer forms, which were in turn followed by cold, and lastly by forms warming to the present. Since the bog lies in the area currently assigned to Cary glaciation, the second cold episode would be the Mankato, when ice stood in New York State to the north.

A second case in point is reported by Deevey (1951, p. 177) who studied pollen profiles from three lakes and a bog in Aroostook County, Maine. He found that the "postglacial" advance of forests was preceded by a "late glacial" time of tundra climate. This would be compatible with the leaching evidence, *i.e.* that Aroostook County was glaciated by Cary ice but that Mankato ice lay to the north, and/or on the highlands of Gaspe, and only tundra invaded this part of Maine.

Likewise, confirmatory evidence comes from the varve chronology, as proposed by Antevs (1953, p. 206) and by Flint (1953, p. 902), who, basing his conclusions partly on bodies of glacio-lacustrine clay, suggests Tazewell, Cary, and Mankato ages for the drifts of New England.

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OROGENESIS AND DEEP CRUSTAL STRUCTURE—ADDITIONAL EVIDENCE FROM SEISMOLOGY[1]

By Hugo Benioff

Abstract

Seismic evidence indicates that the principal orogenic structure responsible for each of the great linear and curvilinear mountain ranges and oceanic trenches is a complex reverse fault. A study of eight regions in which orogenic activity is in progress reveals that these great faults occur in two basic types, here designated oceanic and marginal. Oceanic faults, situated within the oceanic domain, extend from the surface to depths of 550 to 700 km. They exhibit an average dip of 61°. Their elastic strain-rebound characteristics show that these faults are composed of two separate mechanical units—a shallow component extending from the ocean bottom to a depth of roughly 60 km, and a deeper component extending to the 700 km crustal boundary. The marginal faults situated along the continental margins occur in dual and triple forms. The dual faults comprise a shallow member extending from the surface to a depth of approximately 60 km and an intermediate member extending to a depth of 200 to 300 km. The average dip is 33°. The marginal triple form is similar to the dual down to the 300 km level. At this depth the dip changes abruptly to 60° to form a third component extending down to the 650± km crustal boundary. The elastic strain-rebound characteristics of the marginal faults indicate that the components of these structures also move as separate units, although in South America the two lower elements exhibit some evidence for mechanical coupling. In the continental domain the 300 km level thus represents a tectonic discontinuity not as yet revealed by seismic wave-propagation studies but which is apparently the lower boundary of the continents. Since the oceanic faults and the deep components of the marginal faults have the same average dip (61°) it may be assumed that both are fractures in a single, continuous mechanical structure subject to a single stress system. The different average dip (33°) of the marginal intermediate fault components suggests that they occur in a structure mechanically distinct from the deep oceanic and continental layer and that they are activated by a different stress system. A hypothesis offered for the origin of the volcanoes associated with the faults assumes that the source of volcanic energy is heat produced in the fault rocks by the inelastic components of the repeated to-and-fro strains involved in the generation of the sequences of earthquakes and aftershocks.

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[1] Except for the paragraph on the origin of volcanoes and a few minor deletions and additions, this paper was read under the title, *Evidence from seismology on the nature of orogeny*, at the Panel on Orogeny of the Cordilleran Section meeting of the Geological Society of America, Tucson, Arizona, April 11–12, 1952.

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Introduction

The physical science of seismology is based almost entirely on observations made with seismographs and clocks. The principal observed data refer to the origin times of earthquakes, the depths and geographic distribution of their foci, and the amplitudes, frequencies, and propagation characteristics of seismic waves. In the past, seismic contributions to the problem of orogenesis included chiefly information as to locations of regions of tectonic activity and evidence for crustal structure exhibited by the speeds and transmission discontinuities of seismic waves. Publication of the monumental catalog of earthquakes by Gutenberg and Richter (1950) has made it possible to augment these early contributions with knowledge derived from magnitudes, time sequences, and accurate hyperfocal locations. Thus an investigation (Benioff, 1949) of the elastic strain-rebound characteristics and related spatial distributions of foci of the seismic sequences of South America and the Tonga-Kermadec region indicated that the orogenic features associated with these two structures represent surface expressions of great faults extending 650 km in depth and up to 4500 km in length. These preliminary findings were thus at variance with those older concepts in which orogenesis is considered a result of sinking and subsequent rising of weakened portions of a thin (35 km±) crust floating on a viscous or plastic substratum. In the present discussion, studies of an additional number of orogenically active regions are presented. These include the Sunda arc, the Kurile-Kamchatka segment, Mexico and Central America, the New Hebrides, the Philippines, the Bonin-Honshu segment, the Aleutian arc, as well as revisions of the South American and Tonga-Kermadec presentations. Magnitudes, origin times, depths, and geographic locations of the earthquakes and the locations of active volcanoes are taken from Gutenberg and Richter (1950).

Regional Seismic Sequences—Observations

South America

A revised map with composite profiles of the South American earthquake sequences is shown in Figure 1. Epicenters and foci of shallow, intermediate, and deep earthquakes are represented by circles, circular dots, and triangular dots respectively. Active volcanoes are represented by stars. The linear distribution of volcanoes exhibited here is characteristic of all the sequences represented in this study except the Philippines. The oceanic trench

lying approximately parallel to the line of volcanoes is also characteristic. In constructing the composite profiles of these sequences and of the following ones, the plotted horizontal positions of foci represent their perpendicular distances from the volcano lines taken as reference co-ordinates. The region has been divided as shown into northern and southern sections AB and BC, for geometric simplicity of the projections A′B′ and B′C′. The complex South American orogenic fault is a good example of the type which occurs along the margins of continents. Moreover, as the author has pointed out (Benioff, 1949), the strain-rebound characteristics and the distribution of foci indicate that this fault structure is made up of three components having three distinctly different types of tectonic movements. The first component involves the shallow layer of the crust above what may possibly be the Mohorovičić discontinuity. Near the fault this layer is assumed to extend to a depth of approximately 60 km. The second component defines an intermediate layer extending from the bottom of the shallow layer to a depth of 250–300 km. The third component extends from the lower surface of the intermediate layer to a depth of approximately 650 km. In the figure, D′ and E′ are vertically exaggerated profiles taken along the lines D and E of the map to show the spatial relations of the

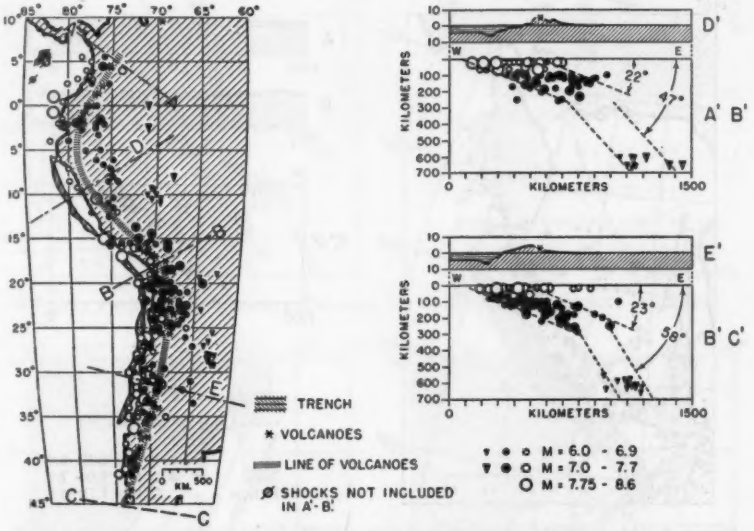

FIGURE 1.—MAP AND COMPOSITE PROFILES, SOUTH AMERICAN EARTHQUAKE SEQUENCES

oceanic trench, the mountain range, and the line of volcanoes (indicated by the star).

In the composite sections A′B′ and B′C′, the vertical and horizontal scales are equal. In the northern (Peru-Ecuador) segment, A′B′, the intermediate component extends to a depth of approximately 250 km with a dip of 22°. In the southern segment (Chile), B′C′, the dip is 23°. Unlike other sequences of this type, South America has no shocks with depths between 300 and 550 km so that the dip of the lower component is not accurately defined. Assuming that there is no horizontal displacement between the intermediate component and the deep component, the fault-zone lines can be extended upward to intersect the intermediate-zone lines as shown, in which case the angles of dip of the deep components of the two segments are 47° and 58° respectively.

Kurile-Kamchatka Segment

Figure 2 shows the orogenic elements of the Kurile-Kamchatka segment. The composite

profile C′D′ represents the portion of the map bounded by CC and DD. A′ and B′ are vertically exaggerated profiles taken along the lines A and B. The Kurile-Kamchatka segment is thus a marginal orogenic fault complex with three components similar to the South American structure. The intermediate component extends to a depth of 300 km and dips under the continent at an angle of 34°. The deep component, with a dip of 58°, appears to have undergone a horizontal displacement of approximately 100 km westward (toward the continent) relative to the intermediate component as indicated by failure of the parallel dashed fault-zone lines to intersect at the 300-km level.

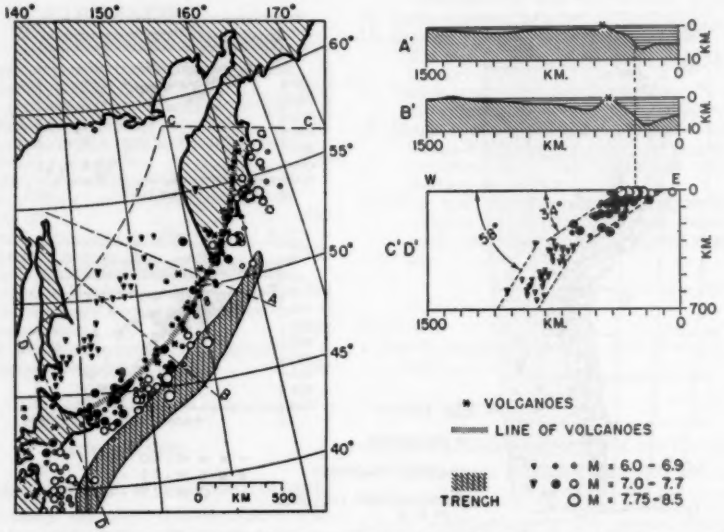

FIGURE 2.—MAP AND COMPOSITE PROFILE, KURILE-KAMCHATKA EARTHQUAKE SEQUENCES

Bonin-Honshu Segment

The region of the Bonin-Honshu segment is shown in Figure 3. The composite profile B′C′ represents the portion of the map bounded by BBCC. A′ is a vertically exaggerated profile taken along the line A of the map. This orogenic complex is actually a part of the Kurile-Kamchatka-Japan structure, but has been separated here to avoid interpretation difficulties arising from distortion of the shallow layer in the vicinity of the Japanese islands. The high shallow layer activity and absence of foci deeper than 150 km along the Nansei Shoto branch to the west, the paucity of shallow earthquake activity in this (Bonin-Honshu) segment, the large horizontal displacements between its two lower components, and its reverse curvature all strongly suggest that the whole fault complex from northern Kamchatka to the southern portion of the Bonin-Honshu segment was once continuous and convex toward the Pacific over its whole extent, as indicated on the left in Figure 4. The present form appears to be a result of an approximately 90-degree counterclockwise bending about a vertical axis in Manchuria, of the two lower components of the original arc relative to the original shallow component now forming the Nansei Shoto branch, as shown on the map on the right in Figure 4. Consequently, in the present Bonin-Honshu segment the shallow component is missing, and the intermediate component extends from the surface to a depth of 400 km. It is thus the deepest continental structure observed in this study. The fault dips westward under the continent at 38°. The deep component is displaced horizontally

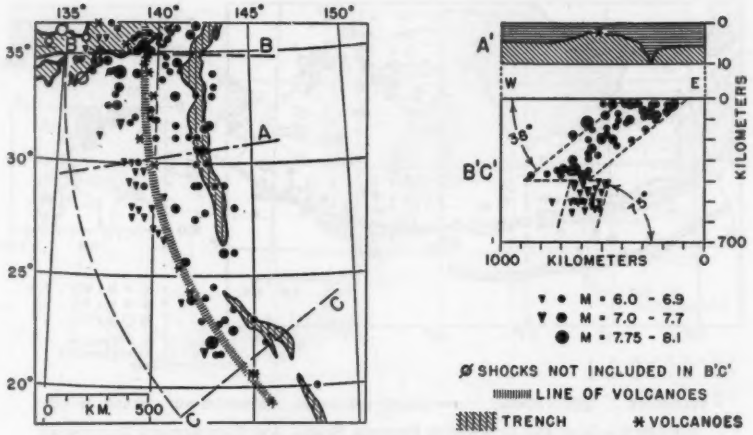

FIGURE 3.—MAP AND COMPOSITE PROFILE, BONIN-HONSHU EARTHQUAKE SEQUENCES

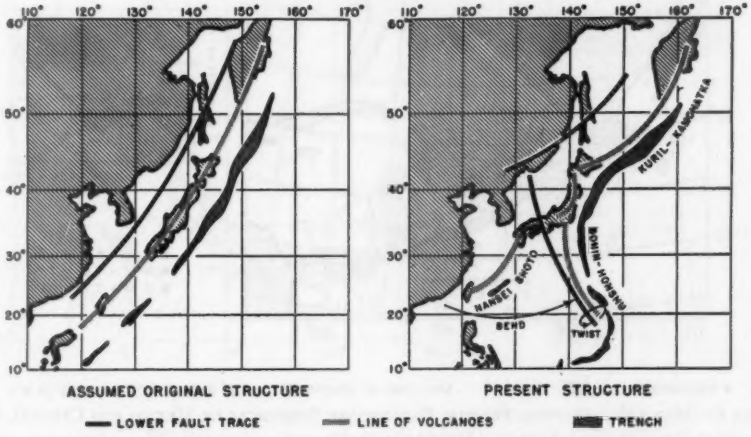

FIGURE 4.—MAPS SHOWING PRESENT AND ASSUMED ORIGINAL CONFIGURATIONS OF BONIN-JAPAN-KAMCHATKA ARC

eastward, away from the continent, some 200 km relative to the intermediate component. This displacement, and the 75-degree dip, the largest of any studied to date, suggest that the forces which originally reversed the curvature of the Bonin-Honshu segment also twisted the deep component fault surface about a horizontal axis, as indicated on the right in Figure 4. It appears therefore that the common component of the couples, responsible for the bend and the twist, acted horizontally eastward along the lower boundary of the lower layer.

Sunda Arc

Figure 5 shows the region of the Sunda arc. The portion of the map represented in the composite profile B′C′ is that bounded by the dashed lines BBCC. This structure is also a triple marginal fault. The intermediate component dips northward under the continent at an angle of 35°. It has a maximum depth of approximately 300 km. The dip of the deep component is 61°. It has been displaced horizontally approximately 200 km southward (away from the continent) relative to the

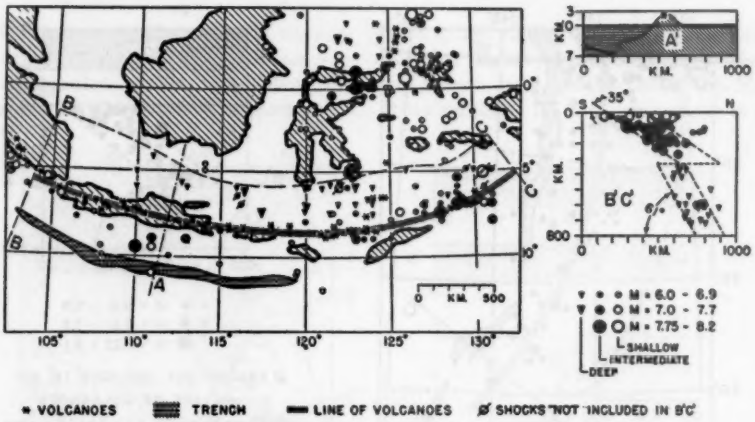

FIGURE 5.—MAP AND COMPOSITE PROFILE, SUNDA ARC EARTHQUAKE SEQUENCES

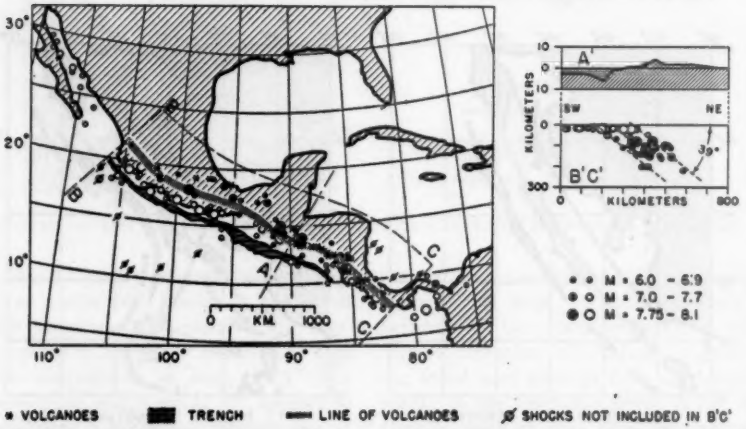

FIGURE 6.—MAP AND COMPOSITE PROFILE, EARTHQUAKE SEQUENCES OF MEXICO AND CENTRAL AMERICA

intermediate component, as indicated by the dashed lines in B′C′.

Mexico and Central America

Figure 6 shows the dual marginal Acapulco-Guatemala fault complex. The intermediate component extends (under the continent) to a depth of only 220 km with a dip of 39°. The shocks northwest of the boundary BB are shallow and presumably represent horizontal movements on the shallow clockwise transcurrent San Andreas fault system. There is no evidence for the existence of a deep component in this region.

Aleutian Arc

Figure 7 refers to the Aleutian arc, also a marginal dual fault. The composite profile B′C′ shows only those features bounded by the dashed lines BBCC of the map. The intermediate component dips northward 28° and extends to a depth of 175 km, thus forming the shallowest continental structure treated in this study, with the possible exception of the Philippines.

New Hebrides Region

Figure 8 shows the region of the New Hebrides fault. This orogenic structure is also a

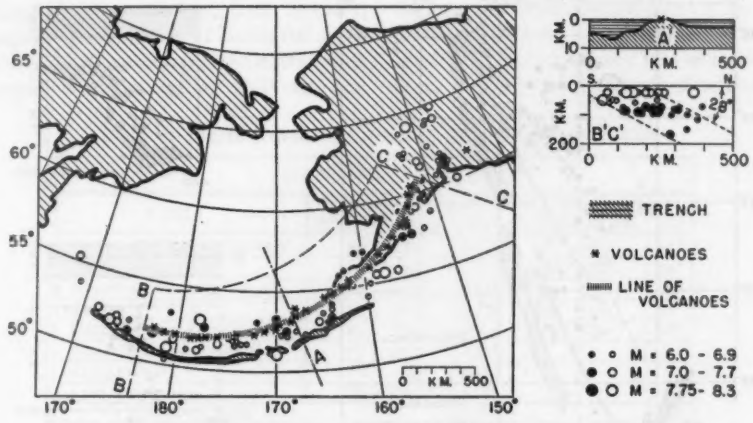

FIGURE 7.—MAP AND COMPOSITE PROFILE, ALEUTIAN EARTHQUAKE SEQUENCES

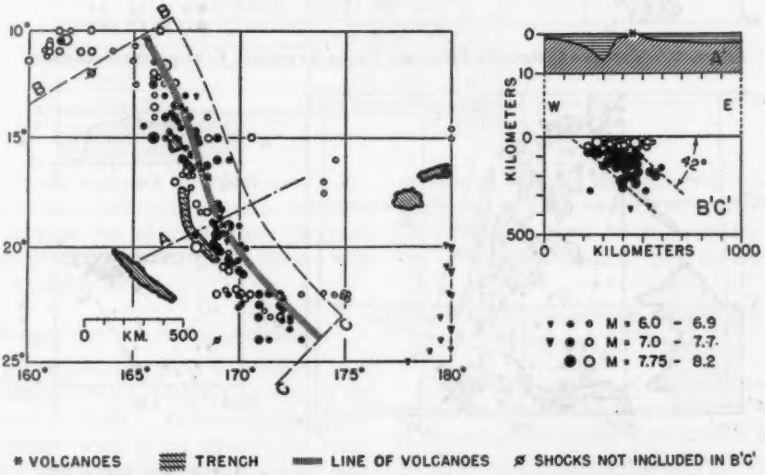

FIGURE 8.—MAP AND COMPOSITE PROFILE, NEW HEBRIDES EARTHQUAKE SEQUENCES

marginal fault principally of the dual type, although the occurrence of two shocks at the 350-km level suggests the possibility of a deeper component of small activity. The intermediate component extends to a depth of 300 km and dips northeast 42°.

Tonga-Kermadec Region

A revised map and composite profile of the Tonga-Kermadec fault complex is shown in Figure 9. In the earlier paper (Benioff, 1949) the author assumed that the structure consisted of a single continuous fault. However, the existence of two lines of volcanoes and the asymmetrical distribution of epicenters are better explained with two faults, even though the strain-rebound characteristics indicate that the two are actuated by a single stress system. The composite profile A′B′ represents the map area AABB associated with the Tonga segment. The vertically exaggerated profile C′ is taken along the line C of the map. The composite profile D′E′ represents the area DDEE of the map, associated with the Kermadec segment. The latter extends well into

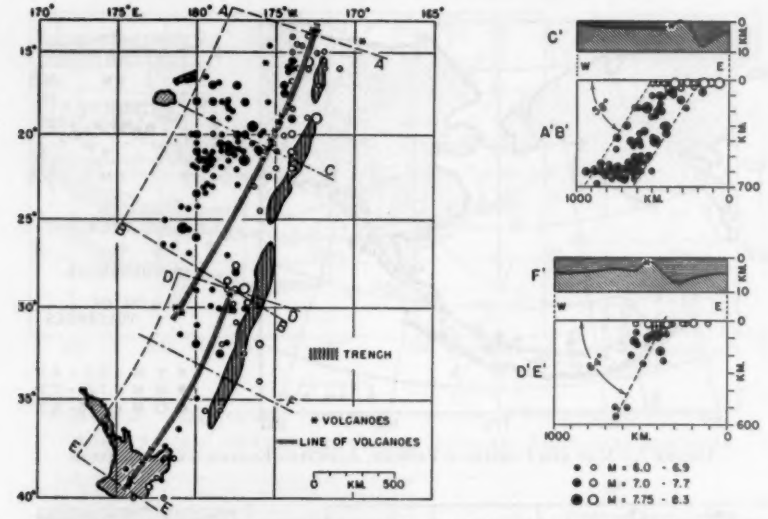

FIGURE 9.—MAP AND COMPOSITE PROFILES, TONGA-KERMADEC EARTHQUAKE SEQUENCES

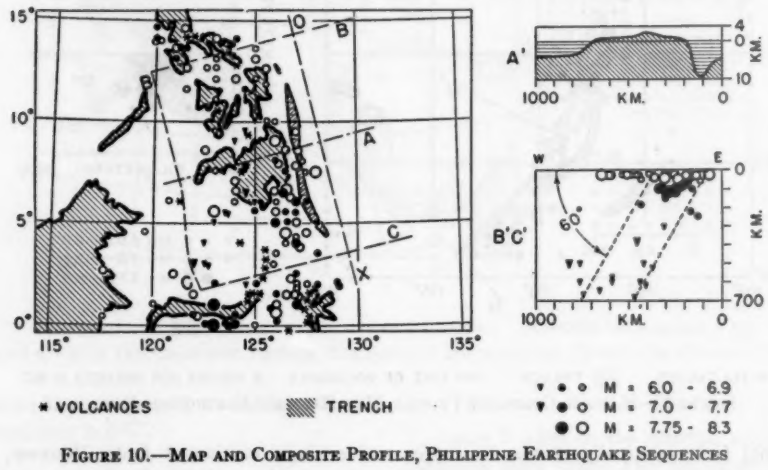

FIGURE 10.—MAP AND COMPOSITE PROFILE, PHILIPPINE EARTHQUAKE SEQUENCES

the region of North Island of New Zealand. The Tonga-Kermadec structure is the principal representative of the oceanic form of orogenic fault complex. The elastic strain-rebound characteristics of these faults (Benioff, 1949) and their composite profiles show that they are of dual type with shallow components extending to a depth of approximately 70 km. The deep component of the Tonga fault dips 59° NW. and extends from 70 km to 650 km in depth. The deep component of the Kermadec fault dips approximately 64° NW. and extends to a depth of 550 km.

Philippine Islands

The region of the Philippine Islands (Fig. 10) is the most disturbed area in this series. The volcanoes exhibit no well-defined line. The projection line B′C′ was constructed with the reference line OX drawn approximately parallel to the strike of the trench. The structure is

considered here an oceanic fault complex, although the concentration of foci in the depth range 70–150 km may represent an intermediate component of corresponding thickness and the mechanism of orogenesis found in the earlier investigation of the Tonga and South American sequences. Thus it appears that in those great linear and curvilinear tectonic structures, now

TABLE 1.—DEPTH RANGES AND AVERAGE DIPS OF OROGENIC FAULTS

MARGINAL FAULTS				
FAULT	INTERMEDIATE COMP.		DEEP COMPONENT	
	DEPTH RANGE, KM.	DIP	DEPTH RANGE, KM.	DIP
PERU - ECUADOR SEGMENT	70 - 250	22°	600 - 650	47°
CHILE SEGMENT	70 - 290	23°	550 - 650	58°
BONIN-HONSHU SEGMENT	0 - 400	38°	400 - 550	75°
SUNDA	70 - 300	35°	300 - 700	61°
KURIL-KAMCHATKA SEGMENT	70 - 300	34°	300 - 650	58°
ACAPULCO - GUATEMALA	70 - 220	39°		
ALEUTIAN	70 - 175	28°		
NEW HEBRIDES	70 - 300	42°		
AVERAGE DIPS		33°		60°

OCEANIC FAULTS		
FAULT	DEPTH RANGE, KM.	DIP
MINDANAO	70 - 700	60°
TONGA	70 - 650	58°
KERMADEC	70 - 550	64°
AVERAGE DIP		61°

having a dip westward of approximately 36°. If this is the correct interpretation, the fault is a marginal one of triple type. The deep component extends to a depth of 700 km with a dip of roughly 60°.

SUMMARY OF FAULT MEASUREMENTS

Depth ranges and dips of the orogenic fault components for each of the regions studied are assembled in Table 1. In the marginal faults the average values of the dips for the intermediate and lowest components are 33° and 60° respectively. In the oceanic faults the average dip of the lower component is 61°. The observational precision (approximately ±50 km in any direction) is not high enough to permit any conclusions regarding the dip of the shallow components nor to differentiate between dip slip and strike slip faulting in their movements.

IMPLICATIONS CONCERNING CRUSTAL STRUCTURE AND OROGENY

Principal Orogenic Mechanism, Reverse Fault

This extended study of tectonically active regions supplements the conclusions regarding seismically active, the principal orogenic mechanism involves a complex reverse fault extending to a depth which varies from 175 km to 720 km depending upon the region.

Shallow and Intermediate Crustal Structure Indicated by Faults

The principal features of the faults and the crustal structure which they define are shown in the generalized section of the crust in Figure 11. The continental marginal structure is indicated on the right, and the oceanic type is shown on the left. The shallow continental layer A, above the Mohorovičić discontinuity M, is approximately 35 km thick as determined by many investigators from wave-propagation studies. It is assumed here to be the same as the layer defined by the shallow earthquake sequences. The strain-rebound characteristics studied so far indicate that, in the vicinity of the faults, shocks belonging to the shallow sequences occur down to a depth of approximately 60 or 70 km in accordance with the classification of shallow, Gutenberg and Richter (1950). Accordingly, in the figure the shallow layer is shown thickened to this extent

in the vicinity of the fault. Presumably this thickening is a result of drag and plastic flow of the strained fault lip. Thus, in part at least, the roots of mountains may be expressions of the oceanic structures the fault extends all the way from this level to the 60-km discontinuity, and presumably to the surface, with a constant average dip of approximately 61°. Since the

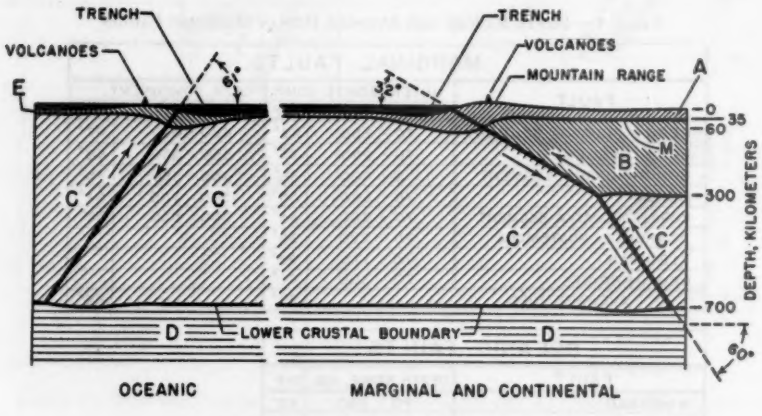

FIGURE 11.—GENERALIZED OCEANIC AND CONTINENTAL DEEP CRUSTAL SECTIONS WITH OROGENIC FAULT TYPES

this fault-generated distortion. In the triple marginal faults the discontinuity in angle of dip at 300 km±, and the marked dissimilarity between the strain-rebound characteristics of the shocks occurring in the fault components above and below this level, are interpreted as evidence for an additional tectonic discontinuity between B and C, as shown in the figure. In the dual structures the lower fault component C is either absent or inactive. The intermediate component with the 32-degree dip occurs in marginal structures, but not in oceanic structures; consequently, the layer B defined by this component is here assumed to be a part of the continental structure. If this assumption is correct, the continents extend downward to an average depth of 280 km, with 175 km and 400 km extreme values.

Deep Crustal Structure Indicated by Faults

In the triple marginal structures the fault dip increases suddenly at the lower boundary of the intermediate component to an average value of approximately 60° and continues down with this value to the 700± km level, below which earthquakes do not occur. In dips of the oceanic faults and the deep components of the continental faults are essentially equal, it is assumed that both types are fractures in a single continuous medium (Fig. 11, C) subject to a single stress sytem. The measured seismic wave speeds near the upper boundaries of the continental layer B and the oceanic layer C are very nearly equal (8.2± km/sec for compressional waves), This suggests that they may be identical in composition. Bullard (1952), on the other hand, cites evidence to the contrary. However, whether or not they are of identical composition, the layers are separated by a tectonic discontinuity and are subject to different stress systems.

RIGIDITY IN DEEP LAYER

Since earthquakes occur down to depths of 550–720 km in both marginal and oceanic faults, it must be assumed that at these depths the crustal rocks can maintain elastic shearing strains of sufficient duration to accumulate the necessary strain energy to generate earthquakes. Moreover, the strain-rebound characteristics of the deep sequence of South America provide evidence for accumulation of elastic

strain without creep or flow over a period of 5½ years at a depth of 550–650 km (Benioff, 1949). The Tonga deep sequence exhibited elastic (recoverable) creep without flow over a period of 15 years down to a depth of 650 km (Benioff, 1949). Thus, for time intervals at least of the order of a decade, the earth's crust in the vicinity of the orogenic faults is a rigid solid down to the 700± km level. Accordingly, this depth limit is here taken as the lower boundary of the solid crust. For geologically long time intervals the crust may, of course, behave as a viscous or plastic substance at these (and shallower) depths. The question as to whether or not the crustal structure, defined in this manner by orogenic faults, extends over the whole earth to regions not now seismically active, is one which cannot be answered by studies of this kind alone. In this connection, Birch's (1951) discovery, derived from calculations of the rate of change of the bulk modulus with pressure, of a substantial departure from physical or chemical homogeneity in the depth range 200–800 km may be additional evidence for the discontinuities and structures derived from these studies.

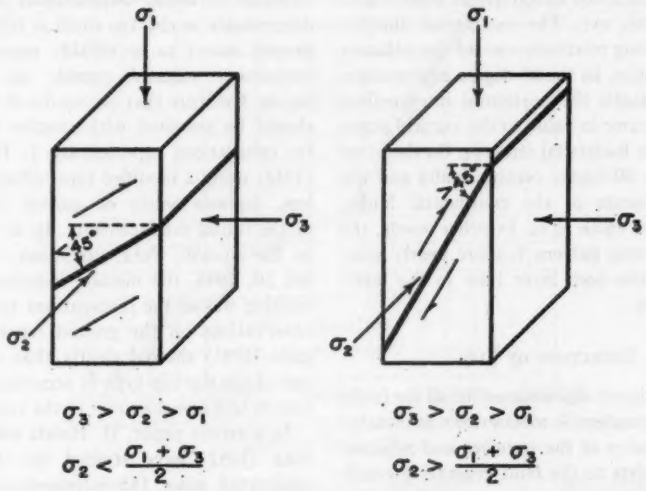

FIGURE 12.—STRESS RELATIONS INDICATED BY FAULT DIPS

OROGENIC FAULT-GENERATING STRESSES

The intermediate components are reverse faults which dip under the continents at an average angle of 33°. This value falls within the range calculated and observed by Hubbert (1951) for faults produced by stress patterns in which the greatest principal stress is a horizontal compression. However, there is reason to believe that the Coulomb-Mohr theory of fracture on which Hubbert bases his reasoning may not be applicable to triaxial stress-conditions of extreme magnitude which must exist at the depths of the intermediate and deep fault components. David Griggs has suggested (Personal communication) that under these extreme conditions the relative value of the intermediate principal stress affects the angle which the plane of fracture makes with the direction of the greatest or least principal stress. Thus, in Figure 12 the block represents portions of a fault structure subjected to the three principal stresses σ_1 (vertical), σ_2 (horizontal parallel to the fracture) and σ_3 (horizontal perpendicular to σ_2). In order that the direction of slip be as indicated by the arrows, the relations of the principal stresses must be such that

$$\sigma_3 > \sigma_2 > \sigma_1$$

If σ_2 is less than the average of σ_1 and σ_3, the angle of dip is less than 45°. If σ_2 is greater than the average of the other two, the angle of dip

is greater than 45°. On this basis it would appear that both the 33-degree and 61-degree faults are produced by stress patterns in which the greatest principal stress is a horizontal compression oriented effectively at right angles to the orogenic axis. The continental margins are thus moving relatively toward the adjacent oceanic domains. In the 33-degree intermediate continental faults the horizontal intermediate stress σ_2 is nearer in value to the vertical stress σ_1 than to the horizontal stress σ_3. On the other hand, in the 60-degree oceanic faults and the deep components of the continental faults, σ_2 is closer in value to σ_3. In other words, the horizontal stress pattern is more nearly symmetrical in the deep layer than in the intermediate layer.

Direction of Slip

The direction of slip assumed for all the faults in which the underside moves down is dictated by the geometry of the trenches and adjacent uplifts. It refers to the fault-generating conditions only. There is no reason to believe that these original conditions necessarily remain unaltered in time, and consequently, once the fault structure has been formed, subsequent stress patterns may change. Later movements may thus involve horizontal slip or even reversals of slip direction. Such alterations may be temporary or permanent. In the latter case, the trench eventually disappears. Perhaps this has been the history of the Pacific coast of North America between Alaska and Lower California and of the Himalayan-Indian arc. A reverse slip may thus be the evidence for decline in the life of an orogenic fault. Instances of reversed slip have been observed by Ritsema (1952, thesis, Univ. Utrecht) for two deep shocks in the Sunda and Philippine regions respectively, and by Koning (1942) for one shock of the Sunda arc. However, the method employed by these authors for measuring the strike and direction of slip is of questionable reliability. It depends upon correct determination from seismograms of the displacement direction of the initial ground motion for each of a rather large number of seismograph stations. Particularly in deep earthquakes, the initial wave motion is nearly always made up of relatively high-frequency components for which most seismographs have inadequate magnification. Moreover, for those critically situated stations to which the initial waves travel in or near the plane of faulting, the direction of initial displacement is either indeterminate or else too small in relation to the ground unrest to be reliably recorded by an instrument situated outside the epicentral region. Evidence that the results of this method should be accepted with caution is provided by calculations reported by J. H. Hodgson (1952) using a modified form which, nevertheless, depends solely on correct observations of the initial displacements. In an application to the Ancash, Peru, earthquake of November 10, 1946, the method indicated that the faulting was of the transcurrent type, whereas observations on the ground reported by Silgado (1951) showed clearly that the faulting was of the dip-slip type in accordance with the known tectonic character of the region.

In a recent paper, H. Honda and A. Masatuka (1952) have studied the direction of motion of some 145 intermediate and deep earthquakes of Japan using observations of a large number of Japanese stations. Since these stations are all in or near the epicentral region of the shocks, the reliability of first-motion determinations is relatively high. Their results indicate clearly that for these shocks the dip of the faults and the direction of slip are in substantial agreement with the assumptions and findings of the present investigation.

Relationship Between Shallow Oceanic and Intermediate Marginal Layers

The continental surface layer of 35 km thickness in which the shallow marginal components occur is generally reduced to approximately 5 km under the oceans insofar as the evidence from wave-propagation studies indicates. On the other hand, the strain-rebound characteristics of the Tonga-Kermadec sequences show that shallow earthquakes occur to a depth of 60 or 70 km in the oceanic structures. Hence, in the vicinity of the oceanic faults we must assume some sort of shallow discontinuity having approximately the same depth as the shallow discontinuity of the continents. Such a discontinuity is therefore shown on the left in Figure 11.

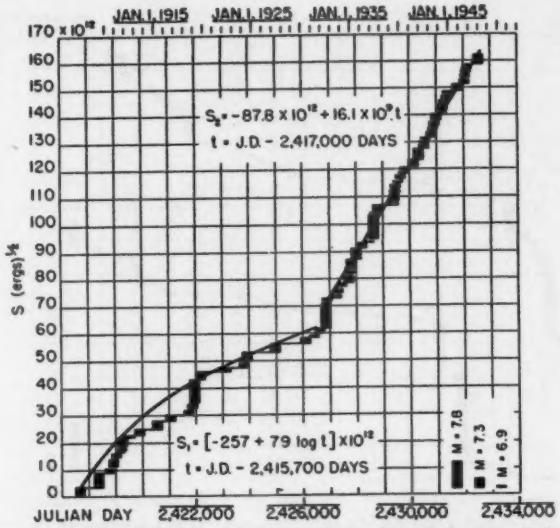

FIGURE 13.—ELASTIC STRAIN-REBOUND CHARACTERISTIC, TONGA-KERMADEC DEEP EARTHQUAKE SEQUENCE
For focal depths 70–680 km

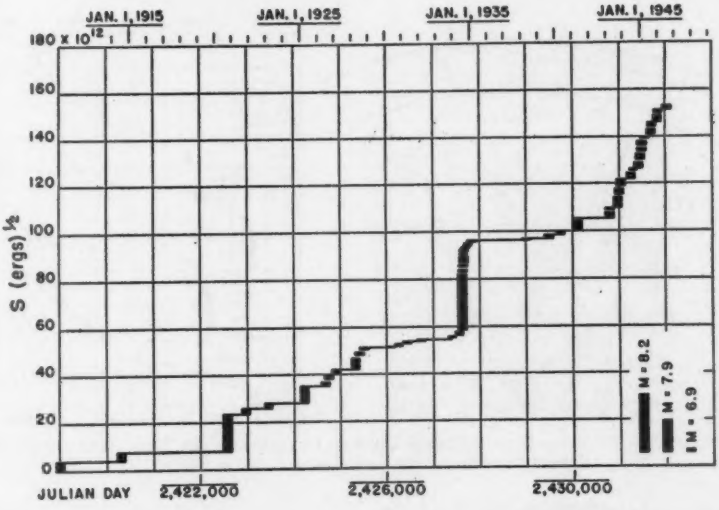

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FIGURE 14.—ELASTIC STRAIN-REBOUND CHARACTERISTIC, NEW HEBRIDES SHALLOW EARTHQUAKE SEQUENCE
$h < 70$ km

As indicated in Figure 11, it is assumed that in general the intermediate layer B terminates at the fault. There is some evidence from the strain-rebound characteristics, however, for the conclusion that in the New Hebrides-Tonga region the intermediate continental layer extends across the intermediate fault to become the shallow layer of the Tonga-Kermadec oceanic fault. The elastic strain-rebound characteristic of the Tonga-Kermadec deep

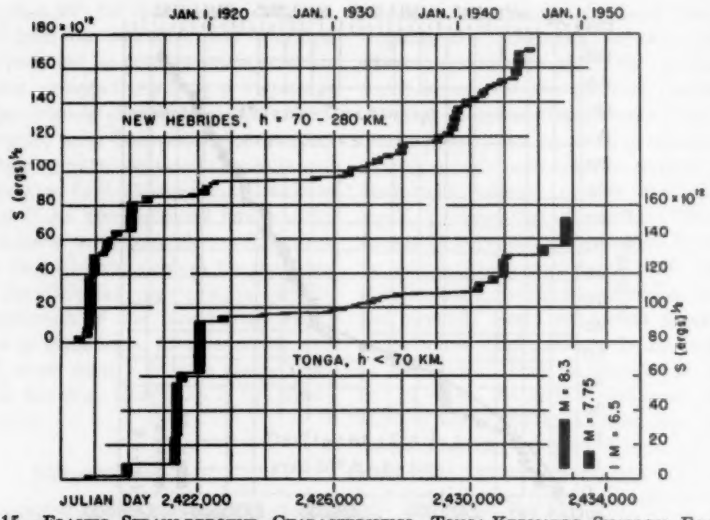

FIGURE 15.—ELASTIC STRAIN-REBOUND CHARACTERISTICS, TONGA-KERMADEC SHALLOW EARTHQUAKE SEQUENCE AND NEW HEBRIDES INTERMEDIATE SEQUENCE
Tonga-Kermadec (h < 70 km); New Hebrides (h = 70–280 km)

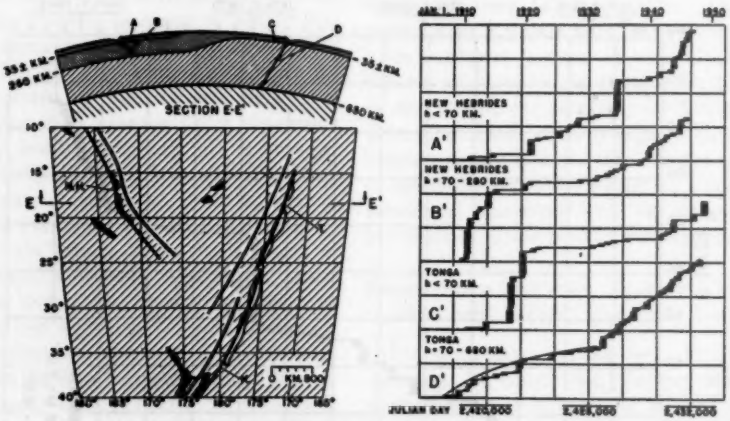

FIGURE 16.—MAP AND ELASTIC STRAIN-REBOUND CHARACTERISTICS, NEW HEBRIDES-TONGA EARTHQUAKE SEQUENCES AND LOCAL CRUSTAL STRUCTURE

components is shown in Figure 13. That for the shallow component of the New Hebrides fault is shown in Figure 14. It is obvious that these two curves are dissimilar and that they must be derived from movements of two distinct mechanical structures. On the other hand, the characteristics of the Tonga-Kermadec shallow sequence and the New Hebrides intermediate sequence (Fig. 15) are strikingly similar, except for a time delay of approximately 2000 days of the former relative to the latter, and both are unlike either the New Hebrides shallow characteristic or the Tonga deep characteristic. It appears, therefore, that the Tonga shallow-fault component and the New Hebrides intermediate component are fractures in a single mechanical structure subject to a single stress system. The situation can best be understood by reference

to Figure 16 in which the left portion is a map and section of the New Hebrides-Tonga region. On the right, the four rebound characteristics are roughly sketched with coinciding time co-ordinates. Counting from the bottom, the first curve—the Tonga deep characteristic—refers to the component D in the section E-E′. The second curve, which is of the Tonga shallow component, refers to the component C of the section. Since these two curves are so very dissimilar, movements in the two involved components are unrelated, and consequently a discontinuity must be drawn between C and D. The third curve is the New Hebrides intermediate characteristic. It refers to the component B in the figure. Since the curves for B and C are so nearly alike we must assume that these two components are members of a single stress system, and consequently the intermediate layer at B is shown continuous with the shallow layer at C. Since the rebound characteristic of the shallow New Hebrides component A (the fourth curve on the right) is unlike that of the deeper component below it, a discontinuity must be drawn between the two. The structure deduced here for the Tonga-New Hebrides region may represent a special configuration of this part of the earth only and may not be valid for other regions in which the structure outlined in Figure 11 must be considered representative.

Orogenesis and Volcanoes

One of the most remarkable features of the orogenic faults discussed in this paper is the linear array of volcanoes lying parallel to the fault strike on the uplifted block. Indeed the line of volcanoes has served as the principal reference co-ordinate for drafting the composite profiles. Volcanoes are clearly one of the manifestations of orogeny. In an attempt to explain this close relationship the writer is offering a hypothesis of the origin of the volcanoes which at least will direct thinking toward this interesting problem. By referring to the composite profiles shown in Figures 1–9 it will be noted that in general the volcano line coincides with the highest elevation of the overhanging fault block and thus appears to be situated along the line of maximum bending of the block. In recent studies the writer (Benioff, 1951) found evidence that the aftershock sequences which follow significant earthquakes are generated by elastic after working or creep recovery of the fault rock. During the interval between principal earthquakes, strains accumulate in the rocks. A portion of this strain is purely elastic, and a portion is time-dependent. When the fault slips to produce the principal earthquake, the purely elastic strain only is involved. The time-dependent or creep strain can be released only relatively slowly in accordance with the creep-recovery characteristics of the rocks. It is this time-dependent elastic recovery which produces the aftershocks. However, only a small portion of the energy stored in the creep-elastic element can be converted into seismic waves in aftershocks. The main portion is liberated as heat. A rough calculation of the energy liberated as heat in a number of aftershock sequences shows that it averages approximately half the amount of the energy liberated as seismic waves in the principal shock. In the orogenic faults on which earthquakes occur repeatedly, the fault rock undergoes a cycle of alternate strain and relief for each earthquake. This repeated to-and-fro bending generates heat in much the same way that quick bending of a wire warms it. The thermal time constant of the great rock masses below the surface is so long that heat generated by these cycles of strain escapes very slowly and consequently accumulates over long intervals of time. If the rock being heated has components with different melting points, there comes a time when one or more will melt. The region of most intense heat generation should coincide with the regions of most intense bending—the highest elevations of the overhanging block and the lowest depths of the neighboring trenches. The low incidence of volcanoes in the trenches may be ascribable to a higher melting point induced by the increased pressure and to the fact that the trench (in the form of a syncline) is not so effective a structure for concentration of the molten rock as the complementary anticline of the uplifted block on the upper side of the fault.

A rough idea of the magnitude of energy released, say per year, by the aftershock se-

quences in a region on one side of the fault can be obtained by taking one fourth of the energy released in the same time by seismic waves in the principal earthquakes. Thus, in the case of South America, the shallow and intermediate earthquake sequences each liberate approximately 4×10^{23} ergs per year. Thus roughly 10^{24} ergs per year is being released in the fault blocks. The writer has no knowledge of the amount of energy per year required to maintain the South American system of volcanoes, and consequently it is not possible to say whether or not the energy requirements are met on this hypothesis. Moreover there must be a large time lag between the liberation of heat in the depths and its appearance in the form of volcano output. Thus the present rate of volcanic energy release should be equated to a phase of seismic-heat generation which occurred long ago, rather than to the present rate.

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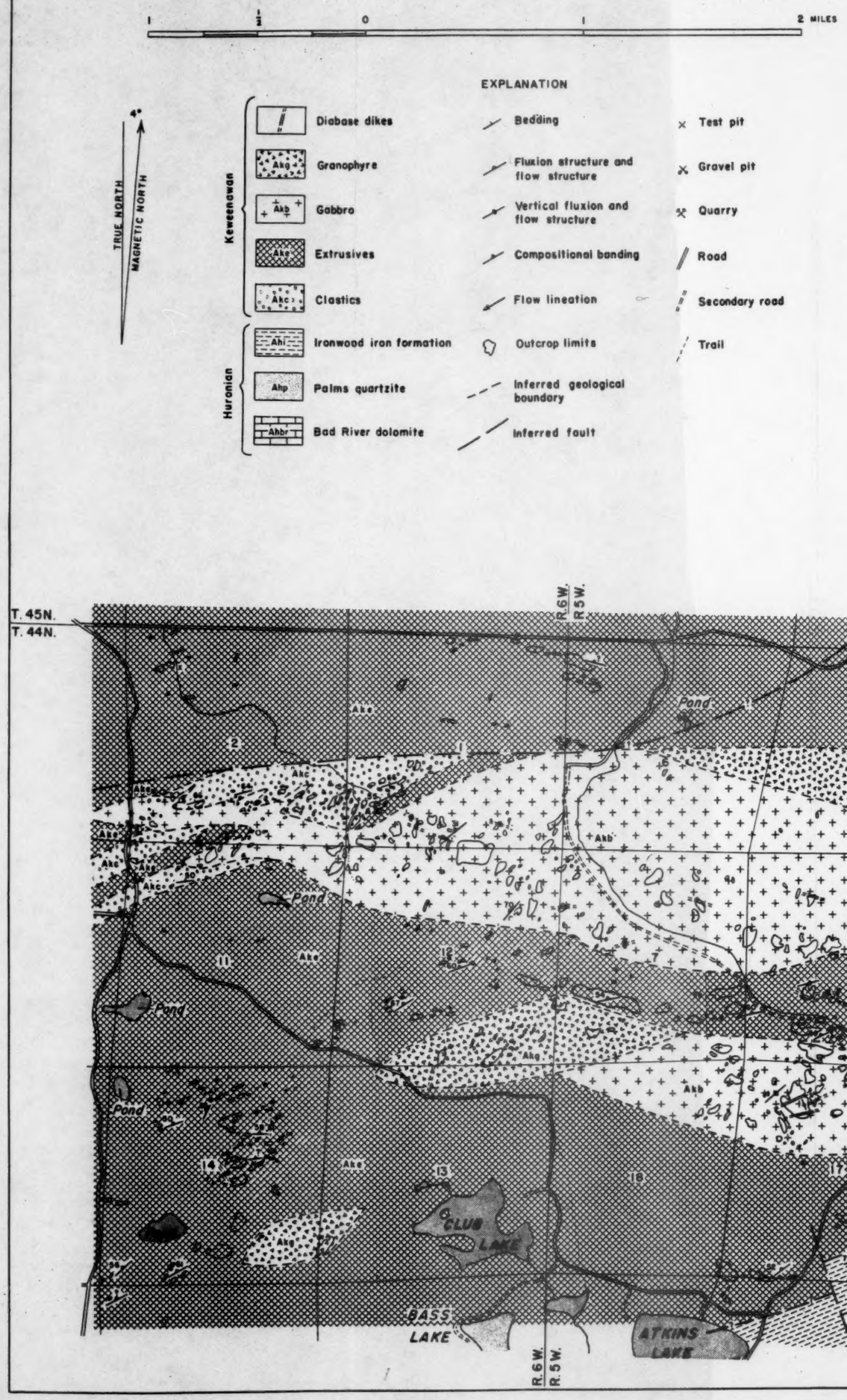
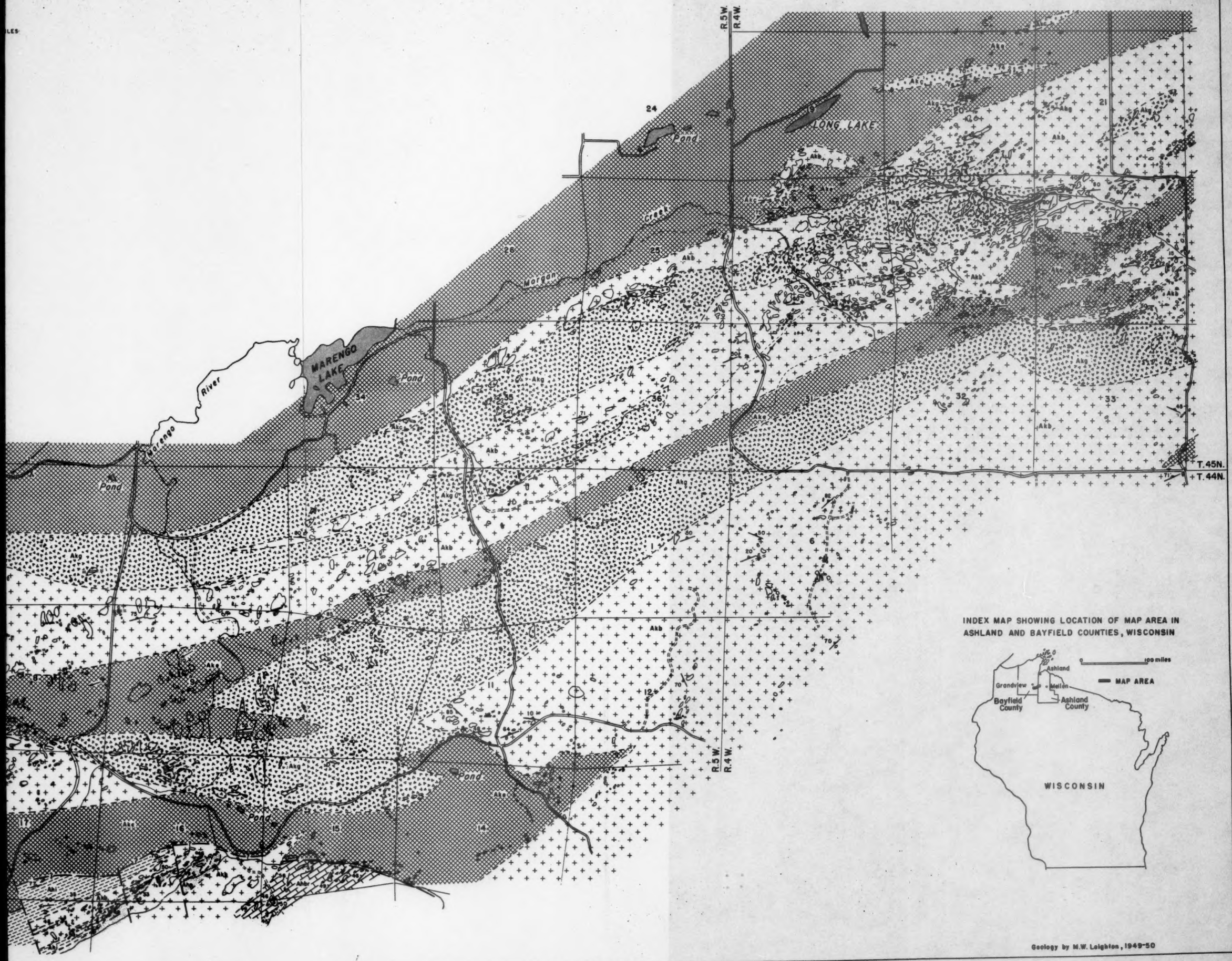


GEOLOGIC MAP OF GABBRO-GRANOPHYRE COMPLEX IN NORTHERN WISCONSIN



GEOLOGIC MAP OF GABBRO-GRANOPHYRE COMPLEX IN NORTHERN WISCONSIN

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PETROGENESIS OF A GABBRO-GRANOPHYRE COMPLEX IN NORTHERN WISCONSIN

By M. W. Leighton

Abstract

Field and laboratory investigations suggest that a granophyre-gabbro complex in northern Wisconsin is similar to other acidic and basic complexes so common throughout the world. The Wisconsin complex consists of two distinct belts of gabbro and granophyre split by a narrow belt of basic extrusives, Keweenawan in age, with flows the oldest, gabbro next, and granophyre the youngest. Still younger are basic dikes that show preferred orientation. The complex is on the steeply dipping south limb of the Lake Superior geosyncline.

The gabbro was intruded between flow units along planes of shear that resulted during formation of the geosyncline. Some gabbroic magma followed thrust faults. Two structures in the gabbro—fluxion structure and rhythmic layering—appear to be unrelated in origin. Fluxion structure is believed to be the result of shear or flow occurring parallel to the walls of the intrusion as the mass crystallized, whereas the layering appears to be the result of crystal settling. A study of mineral compositions in the gabbro reveals simultaneous iron enrichment in coexisting pyroxenes and olivines, and soda enrichment in the plagioclase.

The granophyre is believed to be the result of crystallization of a granitic melt which was intruded following solidification of the gabbro. Criteria for a replacement (metasomatic) origin of the granophyre are inconclusive. It is postulated that the granitic melt originated at depth through either (1) partial remelting of the sial during downwarping of the Lake Superior geosyncline or (2) fractional crystallization. This melt was then squeezed into its present position during further deformation of the geosyncline and followed essentially the same zones of weakness taken by the gabbro magma. The minerals of the granophyre could be the residua of fractional crystallization or the first liquid formed in remelting.

The type of granophyric texture in the red rock (granophyre) is largely dependent upon the anorthite content of the microperthite. In addition, the granophyric texture cannot be the result of eutectic crystallization, for this type of crystallization implies that the melt crystallizes with no resulting change in composition; however, in the red rock, a change in composition is reflected in the anorthite content of the microperthite intergrown with quartz. Hence, the granophyric texture must be due either to simultaneous crystallization without a eutectic or to some secondary process.

Between the gabbro and the granophyre (red rock) is a narrow zone (maximum 200 feet) of "intermediate rock" which exhibits the peculiar mineralogical and textural characters of an alteration zone. Geochemical culminations of FeO, TiO_2, MnO, and P_2O_5 exist in this zone between gabbro and granophyre. In addition, the intermediate zone has been enriched in the O^{18} relative to the O^{16} isotope. The peculiar chemical, mineralogical, and textural features of the "intermediate rock" are the result of contact metasomatism caused by the action of a granophyric melt on gabbro. Also the alteration of the basic flows in the area probably resulted from solutions emanating from the granophyre.

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INTRODUCTION

Statement of the Problem

Numerous bodies of red granitic rocks occur in the Keweenawan of the Lake Superior region. Grout and Schwartz (1939, p. 41) stated that these red granitic rocks are locally known:

"... as 'red rocks', partly because of their rather characteristic, bright, brick-red color (though some may bleach to pink or white), and partly because no other single name covers all the peculiar features

shown by the rocks. Most of them are sodic granites with some graphic intergrowths of quartz and feldspar; granophyre might serve as their name."

In this paper the terms granophyre and red rock are used synonymously.

It is the purpose of this study to determine the origin of these rocks. The area of exposed red rock in Ashland and Bayfield Counties, Wisconsin, especially appeared to warrant a detailed investigation, since there occurs here a belted complex of gabbro and red rock unparalleled in the Lake Superior region. The association, gabbro-red rock or diabase-red rock, is a common one in this region and corresponds to the gabbro-granite associations so common in other parts of the world. As Holmes (1931, p. 241) stated,

"one of the outstanding problems of petrogenesis at the present time is that offered by the remarkable associations of sharply contrasted acidic and basic rocks (*e.g.* granite-gabbro; granophyre-dolerite; pitchstone-tholeiite; and rhyolite-basalt) in igneous complexes such as those of the British Tertiary Province and those of the great lopoliths of Duluth and Sudbury in North America and the Bushveld in South Africa."

Thus, the problem of the red rock is one of petrogenesis involving not only the red rock itself but also the associated gabbro. As Daly (1917, p. 424) pointed out, the problem of the origin of the red rock is identifiable with the general controversy concerning the origin of the world's granites.

Widely divergent hypotheses of origin of granophyric masses are found in the literature. Summaries of many of these hypotheses may be found in articles by Fenner (1926), Hoffman (1930), Shand (1947), and Tilley (1950). In general, the postulated origins, like those for granites, can be divided into two opposing groups: (1) origin through crystallization of a granitic melt and (2) origin through replacement (metasomatism). The applicability of both hypotheses will be reviewed for the Wisconsin gabbro-granophyre complex.

Acknowledgments

The writer expresses his appreciation to Dr. Hans Ramberg and Dr. F. J. Pettijohn for valuable criticism and advice. Dr. S. R. Silverman kindly made the oxygen isotope determinations, and Mr. O. Joensu the spectroscopic analyses of alkalies. The writer is also grateful to Mr. E. F. Bean of the Wisconsin State Geological Survey for access to the Survey files in Madison. The project was financed by an Atomic Energy Commission Pre-Doctoral Fellowship.

Location of the Area

The belted granophyre-gabbro complex stretches across parts of Bayfield and Ashland counties, Wisconsin, between Grand View and Mellen, 20 miles due south of Ashland (Pl. 1, inset). The investigated area is a strip of wooded and hilly country 3–4 miles wide and about 11 miles long extending N. 70° E. from the NE ¼ T. 44 N., R. 6 W., into the SW ¼ T. 45 N., R. 4 W. Physiographically, the belted complex occurs within the Lake Superior Highland Province (Aldrich, 1929). Immediately to the north is the Lake Superior Lowland. The extreme western end of the Gogebic iron range forms the southernmost limit of the area.

Previous Investigations

In the offices of the Wisconsin State Geological Survey are field notes and township reports of the area based on field work carried out in 1915 and 1923. Detailed studies of the gabbro-granophyre complex have not been published.

One of the earliest references to the area was made by Chamberlin (1880, p. 168) who wrote of the 50 mile extension of the Keweenawan Bad River gabbro from the Montreal River into T. 44 N., R. 6 W. Subsequently, Van Hise and Leith (1911) suggested that the gabbro in Wisconsin resembled the Duluth gabbro in every respect. Both were laccolithic, and acidic rocks cutting the coarse gabbros were clearly intrusive. Daly (1914, p. 330) mentioned the Bad River gabbro, indicating the structural relations and form were those of a laccolith or sheet intrusion. He believed that the gabbro and associated red rock were analogous to other gabbro-granite complexes in which the granites were assumed to have a syntectic origin. Grout (1918b) noted that the gabbro in Wisconsin is probably a part of the Duluth lopolith. Hotchkiss (1923) wrote that the gabbro in the present area represented a series of parallel laccolithic

bodies injected into Keweenawan flows, and that the gabbro intrusives were contemporaneous with the Keweenaw thrust fault.

In the same year in a series of unpublished township reports for the Wisconsin Geological and Natural History Survey, H. R. Aldrich briefly described the belted gabbro-granophyre complex in T. 44 N., R. 5 W., T. 45 N., R. 5 W., and T. 45 N., R. 4 W. He suggested that the magma had a tendency to divide into two parts—one being rich in alkalies and silica giving rise to the red rock and the other, rich in lime, magnesia, and iron silicates giving rise to gabbro. The red rock "was an immiscible fraction gathering itself together in masses locally." Aldrich believed that the belted character was the result of a composite of sills or laccoliths, and he pointed out that there appeared to be five belts of gabbro and five of red rock.

Brief references to the Keweenawan and to the present area of investigation are also made in the Wisconsin Geological Survey Bulletin 71 (Aldrich, 1929). Aldrich (1929, p. 74) mentions that the Keweenawan is "sufficiently interesting to warrant a separate volume which is in preparation." However, this volume has not been published. Aldrich's views apropos to the present study regarding the geology of the Keweenawan are mentioned later in this paper.

A summary of the stratigraphy for the area is given by Leith, Lund, and Leith (1935). Other papers have carried brief references indicating the existence of the granophyre-gabbro complex.

Attack of the Problem

Although geologic maps of the area are available in the offices of the Wisconsin State Geological Survey in addition to the map published by Aldrich (1929, Pl. I), the present writer remapped the area to get first-hand knowledge of the geological relationships and to correct the distortion of geologic boundaries which appear on the old maps due to the poor original land survey (Aldrich, 1929, p. 260). Aerial photographs (1:20,000) were used for mapping. Topographic maps are lacking. Field work was carried out during the fall of 1949 and summer of 1950. The field work made several changes in the old maps necessary, but the essential features of the belted complex remain (*compare* Pl. I, Aldrich, 1929, with Pl. 1 in the present paper).

Field Relationships

General Statement

The belted granophyre-gabbro complex lies on the south limb of the Lake Superior geosyncline. Although subordinate folding has produced minor complications, the complex shows an overall dip of approximately 60° N. and a strike varying from east-west in the western end of the area to approximately N. 60° E. in the eastern end.

The complex is essentially composed of three units—flows, gabbros, and red rocks—all Keweenawan. A narrow strip of flows one-sixth to two-thirds of a mile wide divides the gabbro-granophyre complex into two distinct belts (Pl. 1). The granophyre in the northern belt of gabbro-granophyre exhibits several extensions or arms reaching into the gabbro. Isolated patches of red rock are also found in the basic rock. Overlying the northern belt of granophyre is coarse-grained gabbro which, in turn, is overlain by another series of flows.

In contrast to the northern belt, the southern belt of granophyre-gabbro shows red rock overlying gabbro throughout its length. Toward the south-central part of the map area, the gabbro of the southern belt ends abruptly against flows; however, it is found again in two patches—one extending across the northern portions of sec. 16, 17, and 18 in T. 44 N., R. 5 W., and the other occurring in association with the Huronian formations in the southern parts of 15 and 16 and extending into the NE corner of sec. 20 in the same township. The red rock is also discontinuous. Besides the major portion of granophyre in the southern belt, two independent patches occur—one in flows in sec. 13 and 14, T. 44 N., R. 6 W., and the other, a somewhat larger patch, in sec. 12, T. 44 N., R. 6 W., and extending for half a mile east.

Other rock units include the Huronian sediments along the southern boundary in the central part of the map area, and Keweenawan clastics associated with the flows.

Huronian Stratigraphy and Structure

General.—Only brief mention will be made of the Huronian stratigraphy since it has been covered previously by Hotchkiss (1919) and Aldrich (1929). In this area the Huronian consists of the Bad River dolomite, Palms quartzite, and Ironwood iron formation. The Tyler slate found farther east in the Gogebic iron range is apparently absent; however, one large rafted graywacke-slate inclusion believed to represent the Tyler slate was found in a melaphyre flow in the south-central part of sec. 8, T. 44 N., R. 5 W.

Bad River dolomite.—Where exposed in the southern portion of sec. 15, T. 44 N., R. 5 W., the Bad River dolomite is a white, fine-grained, tremolitic marble which weathers to a light yellowish brown. The average strike and dip are N. 40° E., and 45° NW, respectively, but variations between N. 15° E. and N. 70° E. and 31°–55° NW were found. Some folding of the formation has occurred. Aldrich (1929) gave an estimated thickness of 270 feet which the writer regards as extremely conservative.

Palms quartzite.—Stratigraphically overlying the dolomitic marble is the Palms quartzite. A medium- to coarse-grained gabbro intrusion has split the two formations, thus tending to obliterate the unconformable relationship found elsewhere. The Palms shows an upper well-bedded, gray to white vitreous quartzite, and a lower impure quartzite with schistose and slaty interbeds. The folding of the Palms is even more pronounced than that of the Bad River dolomite. Aldrich (1929, p. 262) gives a figure of 360 feet for the thickness of the Palms in this area. Gabbroic intrusions have penetrated some of the quartzite outcrops.

Ironwood iron formation.—Conformably overlying the Palms quartzite is the Ironwood iron formation which is an extremely dense, compact, and well-banded rock. Some outcrops show that the iron formation has been intensely folded. A bedding fault along which the gabbro has been intruded has divided the iron formation into two belts (three in one place). In addition, two and possibly three transverse faults accompanied by inwelling gabbro have offset the belts along their strike. These belts trend approximately N. 45°–50° E.

Whereas Plate 1 suggests that the Ironwood formation is nearly 1150 feet thick, Aldrich (1929, p. 93) stated that the total thickness is only 650 feet. Folding or faulting, perhaps accompanied by intrusion, could account for the apparent local thickening. In addition, the northern outcrop mapped as iron formation may actually be a magnetite-rich phase of the Tyler slate (Aldrich 1929, p. 105–106).

Keweenawan Stratigraphy and Structure

General.—Previous investigations have revealed that the break between the Huronian and Keweenawan is either one of faulting or erosional unconformity or both. Aldrich's views of the importance of faulting (1929, p. 122) have definite support in the present area. Here, along the Marengo River in sec. 15, T. 44 N., R. 5 W., the Bad River dolomite ends abruptly against badly sheared flows. The Palms quartzite and Ironwood iron formation also end abruptly in sec. 16 of the same township. To the west, however, a later gabbro intrusion and lack of outcrops make the position of the fault (the Crystal Lake-Atkins Lake fault) less certain.

The Keweenawan consists of flows, clastics, intrusive gabbros, and the "red rocks" or granophyres. According to Aldrich (1929, p. 267, 274), the flows in the area are middle Keweenawan.

Flows.—Outcrops of flows are not numerous and are generally small. Thus, differentiation into individual flow units is almost impossible. In mapping the flows, the general classification applied to other extrusives in the Lake Superior region was followed (Butler and Burbank, 1929; Sandberg, 1938). Since only brief descriptions of the Keweenawan flows are presented here, the reader is referred to articles by Cornwall (1951) and Broderick (1935) for more detailed work in addition to the above-mentioned articles by Butler and Burbank and by Sandberg.

Basic flows predominate. The acid flows are confined almost entirely to the series of flows overlying the northern gabbro-granophyre belt and are particularly abundant in the northeastern corner of the map area. The flow units show a mean (30 measurements) orientation of N. 70° E. with a dip of 63° N. Of the

30 measurements, 66 per cent show a strike between N. 60° E. and E–W. The direction of flow could not be ascertained, but elsewhere Hotchkiss (1923) has demonstrated the likelihood that the source was to the north along the axis of the Lake Superior geosyncline. Most flows are badly fractured and, near gabbros, show biotite porphyroblasts, indicating some recrystallization. Several outcrops of flows have been brecciated by aplite stringers and dikes or by gabbroic apophyses.

Clastics.—Interbedded with the extrusives are rocks of clastic origin. Some are breccias with angular volcanic fragments, whereas others consist of more rounded volcanic fragments imbedded in a very fine matrix. In many cases it is difficult to tell whether these rocks represent conglomerates or agglomerates because (1) metamorphism has recrystallized them into a dense, compact rock obliterating any possible sand grains, and (2) most outcrops are covered with fungus, obscuring the relations.

One remarkable sequence of these rocks is found on Davis Hill in the extreme western end of the area in sec. 1, 2, and 11, T. 44 N., R. 6 W. Here, the thickness is estimated at nearly 1200 feet. This has previously been called the Great Conglomerate by Aldrich (1929, p. 111, 274). It consists of rounded, pebble- to boulder-sized material in a finely crystalline matrix. The strike and dip of the Great Conglomerate is about N. 70° E., 55°–65° N. Elsewhere, similar clastic rocks are more angular in places approaching breccias. Such rocks are found in sec. 34, T. 45 N., R. 5 W. and sec. 20 and 24 in T. 45 N., R. 4 W.

Gabbro.—Gabbro intrudes the Great Conglomerate of Davis Hill. The main shoot of gabbro cross-cuts the conglomerate, but several arms project from the gabbro into the conglomerate parallel to the bedding. One chilled contact against the conglomerate was observed. Since the Great Conglomerate has been called both middle Keweenawan (Butler and Burbank, 1929, p. 18) and upper Keweenawan (Aldrich, 1929, p. 274) the gabbro intrusion cannot be dated more accurately than post-lower Keweenawan. However, Leith *et al.* (1935, p. 11) refer the gabbro to the middle Keweenawan.

The gabbro is a dark-gray, medium- to coarse-grained rock throughout much of its outcrop area. Some portions, however, become medium- and even fine-grained, as, for example, in sec. 28 and 29, T. 45 N., R. 4 W., where outcrops indicate an abundance of included flows, or in those areas of small intrusions such as in the iron formation. In fact, it is difficult to differentiate between the finer-grained gabbros and the coarse-grained flows at these places. Basic pegmatites composed of extremely large pyroxene and plagioclase crystals occur as irregular segregations in the coarser varieties of gabbros and are especially common in the western half of the northern gabbro-granophyre belt. Some porphyritic gabbro phases outcrop in the south of sec. 19, T. 45 N., R. 4 W. and in the SE ¼ sec. 11, T. 44 N., R. 5 W.

In most outcrops the gabbro is massive; however, at some places structure is well exhibited. This structure is of two types, that due to planar orientation of plagioclase laths and that due to variations in mineral proportions which result in banding. Grout (1918a, p. 439, 446) has termed these two types of structures fluxion structure and banding, respectively. Wager and Deer (1939, p. 37) use igneous lamination for the parallel arrangement of platy minerals and rhythmic layering for the banding.

The fluxion structure or igneous lamination has a rather definite orientation. Seventeen of the 34 measurements taken ranged in the acute angle between N. 80° W. and N. 60° E. All but one of the 34 showed a north-dipping component. The dip of the 17 measurements N. 80° W. to N. 60° E. averages 55°–60° N. Field observations suggest that fluxion structure was best developed in the southern belt of gabbro. Here, the ratio of plagioclase to pyroxene also seems greater. In fact, some of the gabbro of the south belt approaches anorthosite.

Banding or rhythmic layering was noted in only three outcrops. These layers show varying proportions of plagioclase and pyroxene and measure from fractions of an inch to several feet in thickness. Their orientation averages N. 70° E., 39° N. The dip of the banding is nearly 20° less than the average dip of the fluxion structure.

The belts of gabbro appear to belong to separate intrusions. However, Aldrich (1929, Pl. I) and Leith *et al.* (1935) show that these

gabbro belts coalesce with a larger gabbro body immediately to the east. This means that the various gabbro belts are but offshoots from the same gabbro intrusive.

Intermediate rocks.—In many places between the gabbro and granophyre an intermediate zone consists of a peculiar rock which appears to be a mixture of the two rock types. The width of this zone varies from zero to nearly 200 feet. Toward the gabbro, the intermediate zone resembles the gabbro, but closer to the red rock it looks more like red rock. The perfect gradation from the intermediate zone into the gabbro is well shown in several outcrops. There are no sharp breaks; however, the intermediate zone in the immediate vicinity of the red rock is not exposed, and thus the relations at the "contact" are obscure. The intermediate zone is characteristic of the northern gabbro-granophyre belt. If outcrops were present along the contact between the gabbro and granophyre in the southern belt, undoubtedly the intermediate zone would be found there, too. A small intermediate zone, only a few feet wide, occurs between the red rock and flows in sec. 9, T. 44 N., R. 5 W.

Granophyre.—The red rock or granophyre is a massive rock, brick-red in outcrop and composed of quartz and red feldspar with mafics (biotite, chlorite, and magnetite) amounting to less than 5 per cent in most places. A few outcrops appear to be syenitic, but examination with a hand lens shows the quartz intimately and abundantly intergrown with the acid feldspar. Generally the texture is equigranular and fine- to medium-grained; however, in both granophyre belts the grain size reaches a maximum in the east. Here, the outcrops are more like true medium-grained granites than elsewhere in the area. Some porphyritic phases of the red rock occur particularly toward the western parts of the granophyre belts.

The variability within one belt is as great as the variability between both granophyre belts, so that a comparison between the two is difficult. However, the ferromagnesian content is highest in the extreme east end in the south belt. In contrast to Aldrich (1929, Pl. I), the writer maps this southern belt continuous from the SW corner of sec. 31, T. 45 N., R. 4 W. to the NE corner of sec. 33, same township. The finding of additional red rock outcrops in this particular area substantiates this continuity.

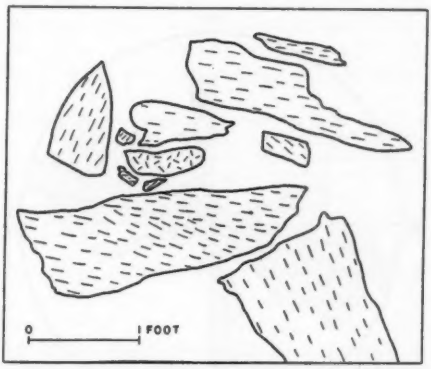


FIGURE 1.—SKETCH OF ROTATED BLOCKS OF GABBRO IN GRANOPHYRE

Fluxion structure in gabbroic fragments indicated by dashed lines.

Such features as red rock dikes cutting gabbros and flows, the presence of a breccia zone of gabbro blocks in a matrix of red rock, and existence of miarolitic cavities in the red rock are of special interest. The dikes and apophyses of red rock commonly wedge out or show varying dip and strike. A small intermediate zone up to several feet wide is found around many of these dikes cutting the gabbro. Chilled contacts are absent in most places, but the break between red rock dikes and the intermediate rock is abrupt. On the other hand, the intermediate rocks grade into gabbro. The contact of one such dike with the intermediate rock is sharply defined owing to the sudden change in both texture and in the amount of ferromagnesians. Similar contact relations have been reported by Krokström (1932, p. 295) for the Breven dolerite dike.

In the NE ¼ sec. 29, T. 45 N., R. 4 W. just south of the narrow projection of gabbro and along Morgan Creek gorge, a breccia of angular gabbro blocks is exposed in red rock. The entire breccia zone is not over a few hundred feet square. The blocks of gabbro, ranging in size from a few inches up to 3 feet, exhibit fluxion structure. This structure shows a different orientation in each block (Fig. 1)

indicating mechanical rotation. Stratigraphically above the breccia is a gabbro with prominent fluxion structure showing orientation of N. 60°–70° E. and 70°–80° N. In a direction

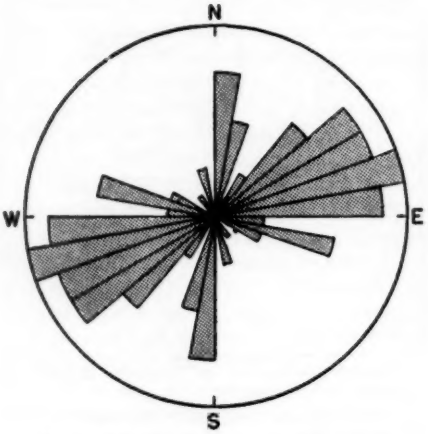


FIGURE 2.—PREFERRED ORIENTATION OF BASIC DIKES
Strike measurements of 62 basic dikes. Radius of circle equals 8 dikes.

from the granophyre toward the breccia zone, the grain size of the red rock decreases. In the breccia zone itself, the red rock becomes fine-grained and contains up to 15 per cent ferromagnesian minerals. The blocks of gabbro are angular, and all are sharply defined.

Dikes.—Numerous dikes of fine-grained diabase, basalt, and porphyritic basalt cut sharply through all Keweenawan rocks. They range from a few inches in width to nearly 20 feet. Chilled contacts are numerous. These dikes have been plotted in Plate 1, and a diagram illustrating their orientation appears in Figure 2. They reveal a preferred orientation in two directions—maximum orientation occurring at approximately N. 5° E. and at N. 70°–75° E. The latter is the more pronounced. From the map, the stronger maximum is found to be due to dike orientation in the western two-thirds of the area; the weaker maximum results from the tendency toward N.-S. orientation in the eastern one-third. Of 53 available dip measurements, approximately 80 per cent (or 42 dips) were steeper than 60°. Dips to both north and south for the E.-W. oriented dikes were recorded. Most of the measurements (66 per cent) possessed a south-dipping component.

Post-dike deformation was recorded in several outcrops. The off-sets were seldom more

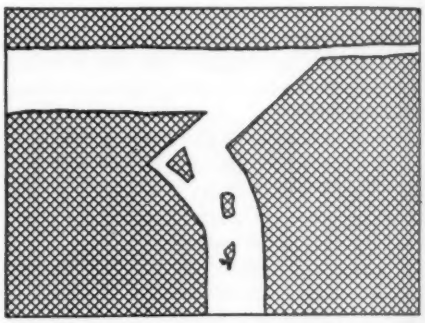


FIGURE 3.—COMPOSITE DIKE
Red rock (white) follows middle of diabase dike (lined) which cuts gabbro (not shown). Scale: Width of figure 1¼ feet.

than a few feet. One basic dike showed a tendency to follow a joint pattern as it zig-zagged its way across the outcrop. Gabbro xenoliths were noted in two dikes.

Acidic dikes are extremely variable in texture. Red rock dikes similar to the main granophyric bodies are the most common. Felsite porphyries are also present, and in one locality a granitic pegmatite was found. Most dikes are massive, but flow banding is particularly well developed in one felsite porphyry dike which intersects both red rock and gabbro in the south of sec. 25, T. 45 N., R. 5 W. These acidic dikes show no preferred orientation.

Several red rock dikes possess fragments of medium- to coarse-grained gabbro and fine-grained diabase. Two composite dikes were found, both consisting of red rock and very fine-grained diabase cutting gabbro. Figure 3 illustrates the relations of one of these composite dikes. The fine-grained diabase of this dike showed a chilled contact with the gabbro. The red rock dike with parallel walls follows the central portion of the basic dike but does not exhibit chilled borders.

The acidic and basic dikes have been particularly useful in establishing age relations.

The gabbros are known to be intrusive into the flows and conglomerates because gabbroic stringers occur in flows, and the chilled contact of a gabbroic dike cuts the conglomerate of Davis Hill. In addition, some flow fragments have been included in gabbro. No gabbroic stringers were found in red rock, but red rock apophyses are extremely common in gabbro and are present in flows in a few places. The age relations of the various dikes are yet to be determined. Composite dikes of red rock and diabase suggest that the red rock is later than the diabase. On the other hand, diabase dikes cut the main red rock masses in a few places, including the breccia zone along Morgan Creek. This anomaly can be resolved by assigning a longer time interval for the formation of diabase dikes than for the red rock. That is, diabase intrusion began prior to and ended after the formation of red rock. Whether or not the diabase dikes are younger than the felsite porphyry dike found cutting both granophyre and gabbro is a problem still unsolved.

In summary the chronological sequence for the various Keweenawan "igneous" rocks is (1) basic and acid flows—early flows were basic; (2) gabbro; (3) diabase dikes (a few); (4) red rock; and (5) diabase dikes (most of them) and felsite porphyry dikes.

Structural Interpretation

Development of monoclinal structure.—Two ideas have been advanced explaining the development of the monoclinal structure *i.e.*, the south limb of the Lake Superior geosyncline. A summary of these ideas was given by Butler and Burbank (1929, p. 50). The first concept, advanced by Van Hise and Leith (1911) and supported by Thwaites (1935, p. 225), states that compressive stresses were dominant and that a thrust from the south against the shield area to the north effected the folding of the geosyncline. On the other hand, Hotchkiss (1923) and Aldrich (1929, p. 155) believe that the geosyncline formed through foundering of the roof of a batholith from which the Keweenawan flows had come. In either case the authors agree that the gradual lessening of dips in flows from lower to higher horizons in the Keweenawan indicates that deformation proceeded during extrusion.

Oblique folds.—Differential tilting during deformation is recorded by the fact that, in the northeastern part of the Gogebic range, the flows dip at a steeper angle than in the western section (Aldrich, 1929, p. 115). As a result of the differential tilting, Aldrich states:

> "warping under torsional stresses became effective .... One principal result of the torsional stresses was the development of two subordinate, oblique, structures on the south limb of the ultimate structure, a syncline striking slightly east of north (15°–30°) and a monocline or anticline immediately to the west and striking 60°–70° west of north."

According to Aldrich (1929, p. 119) the oblique synclinal fold occurs in the southwest of T. 44 N., R. 4 W., and the anticline, farther west at Atkins Lake. Since the folding involved both the Huronian and Keweenawan, Aldrich suggests that the folding took place in post-Keweenawan time.

From the data supplied by Aldrich, a projection of the synclinal structure would cut the present area in the northeastern parts (probably most of the mapped areas in T. 45 N., R. 4 W.). Slight synclinal warping toward the central part of the area is also suggested by Plate 1. The anticlinal structure cuts across the southwestern corner of the map area. Atkins Lake marks the position of the anticline. Thus, a large part of the mapped area lies on a limb between an anticlinal and a synclinal structure.

Thrust faults.—In addition to the torsional warping there was some lateral compression during the development of the major Lake Superior syncline. The compression was relieved by differential updip movement of upper beds over underlying ones. Aldrich (1929, p. 132) believes that the development of the Keweenaw fault and related Crystal Lake-Atkins Lake fault, which passes through the south of the map area along the Marengo River, resulted from these compressional forces.

Aldrich also (1929, p. 125) postulated an additional thrust fault across the north of T. 44 N., R. 6 W., which was a result of the same forces. The conglomerate and traps of Davis Hill are "in structural convergence with

the traps to the north" as revealed by magnetic data and other field observations. The northern flow series strike E.-W. and stand on edge; whereas, the conglomerate and southern flow series strike about N. 60° E. with a more gentle northern dip. Aldrich shows that the reason for the convergence is the presence of the Lake Owen thrust fault. Plate 1 shows the approximate location of this fault as it cuts through sec. 1 and 2, T. 44 N., R. 6 W. Its position farther east is more arbitrary.

Gabbro intrusion and structure.—The gabbro intrusion has followed planes of weakness developed during the formation of the Lake Superior syncline. Undoubtedly, the torsional warping resulted in differential movement, thus producing two types of structurally weak zones—those due to oblique faulting and those due to slippage between flows. Both structures exerted control on the gabbro in the present area, but slippage between flows was probably the more influential. Only the southern patch of gabbro can be directly related to an oblique fault, the Atkins Lake-Crystal Lake thrust. The remaining gabbro belts are essentially parallel to the flow series except for some cross-cutting in the western end of the map area.

The parallel attitude of the gabbro and flow belts is reflected by the fluxion structure in the gabbro as well as by the general trend of the belts. The strike of the fluxion structure generally ranges from N. 80° W. to N. 60° E. with an average dip of 55°–60°N. This is in good agreement with the average orientation of flows (N. 71° E., 63° N).

Since deformation had begun before intrusion and the flows were to some extent already tilted to the north, crystal settling could hardly give rise to fluxion structure (or igneous lamination). That is, crystal settling would have resulted in practically horizontal orientation of plagioclase plates—this orientation being at wide variance with that of the tilted flows. Since there is no variation in orientation, the fluxion structure is believed to be due to orientation produced by friction during magmatic movement. This movement was controlled by the planes of injection, or, in other words, the movement was parallel to the layering of the extrusive rocks of the region.

Whether the orientation was produced by movement of the basic magma as a result of convection currents (Grout, 1918a; 1918c; 1920; Wager and Deer, 1939, p. 272) or as the consequence of primary injection of a melt already begun crystallizing is a matter of speculation. Ritchey and Thomas (1930, p. 98) state that fluxion gabbros may have intruded while partly solid. The parallel arrangement of platy minerals cannot be due to metamorphic shear since evidence of rupturing or strain in these rocks is generally absent.

The writer agrees with Daly (1928, p. 728) that the parallel arrangement must be a primary feature resulting from flow or shear as the rock crystallized. The views of Balk (1927) with regard to the funnel-shaped Peekskill norite are also applicable here. He suggested that the "flaser" structure, similar to the fluxion structure, was due to frictional adjustment during movement of the mass after the first crystals had formed.

Another internal structure of the gabbro, the rhythmic banding, is believed to have an origin different from that of the fluxion structure. The average dip of the banding is 15°–20° less than the average dip of the fluxion structure, and the two structures were not found together. The recorded differences in dip are believed to be significant even though only three outcrops revealed the rhythmic banding.

One possible explanation for these differences is that the rhythmic banding was formed through differential crystal settling. The crystal accumulations controlled by gravity would lie more nearly horizontal than the fluxion structure which was controlled by movement parallel to the tilted walls of the intrusion. Both Bowen (1928, p. 169) and Van der Berg (1946) emphasize that the banding is a sedimentation structure. Possible mechanisms producing the rhythmical nature of the bands have been expressed previously by Grout (1918a), Bowen (1928, p. 168–169), Hess (1938), and Wager and Deer (1939, p. 272).

Time of tilting.—That tilting preceded intrusion is based on the following: (1) intrusion has followed and cut across the Keweenaw thrust fault formed during tilting; (2) dips of the fluxion structure and rhythmic banding are different, suggesting tilting prior to intrusion as a cause of this difference; (3) shearing

which would accompany tilting of the gabbro mass is absent in the gabbro but present in the flows; (4) flows which preceded intrusion show lessening of dips in higher horizons, indicating tilting during extrusion; and (5) the lack of outcrops showing banding suggests a structure unfavorable to their production, *i.e.*, one in which tilting had destroyed a horizontal floor favorable for crystal accumulation.

In addition, some tilting followed intrusion of the gabbro. This is shown by the present angle of dip of the banding, by the slight off-set of dikes cutting the intrusion, indicating post-intrusion movement, and by the orientation of dikes as described below.

Orientation of dikes.—Reference to Figure 2 shows the pronounced orientation of dikes. This orientation is explained diagrammatically in Figure 4. The N.-S. to N. 10° E. maximum (Fig. 2) may represent an orientation of dikes due to the influence of the oblique synclinal structure mentioned above. These dikes have apparently followed tension fractures developed as a result of synclinal flexuring. The major maximum (Fig. 2) occurring at about N. 70°–75° E. is due possibly to one or more of the following related structural features: (1) tension fractures developed during northward tilting; (2) breaks developed as a result of shear parallel to the layered structure of the flows of fluxion structure of the gabbro; (3) tension fractures developed as a result of the oblique anticlinal structure mentioned above. It is felt that (1) and (2) would explain the observed orientation more correctly than (3). However, the two dikes in sec. 14, T. 44 N., R. 6 W., may have resulted from this anticlinal flexuring, inasmuch as they lie close to the axis of this oblique anticline.

Red rock and structure.—The red rock is nearly structureless except for joints. Some outcrops show as many as three, four, or even five sets of joints making their interpretation nearly impossible. In relation to the overall structural picture the red rock occurs essentially parallel to the gabbro and flows. Contacts with the gabbros and flows are generally not well exposed, but for the most part they are steep *e.g.*, contacts are vertical along the Marengo River where red rock cuts the flows.

The granophyre is found between the two oblique structures. It ends to the east, where the synclinal structure occurs, and dies out to the west near the anticlinal structure. The position of the granophyre with regard to the structure is believed to be significant and bears on its origin.

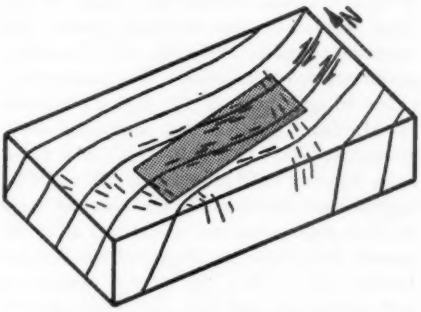


FIGURE 4.—BLOCK DIAGRAM SHOWING STRUCTURAL ELEMENTS

Dashes represent dike orientations. Layered elements represent series of flows and units of shear in gabbro. Stippled portion covers approximate area of investigation.

PETROGRAPHIC DATA

Huronian Formations

The Huronian formations are considerably metamorphosed. A large part of this metamorphism can be attributed to the Keweenawan intrusives; *e.g.*, Aldrich (1929, p. 81, 87, 95) found a direct correlation between the amount of recrystallization and distance from the intrusives.

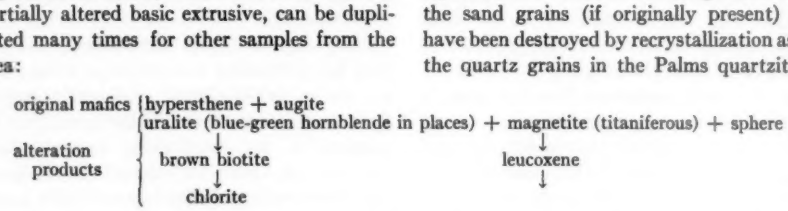
In the present area metamorphism of the Bad River dolomite has produced a light-gray to white, well-laminated, dolomitic marble. Recrystallization of the Palms quartzite has resulted in an interlocking crystalloblastic texture. All traces of original clastic grains have been destroyed. The Ironwood iron formation also shows considerable recrystallization and other effects of metamorphism. Under the microscope, this formation exhibits a crystalloblastic texture with three minerals predominant—quartz, magnetite, and grunerite. These three minerals, occurring in varying propor-

tions, produce the banding observed in the iron formation.

Keweenawan Extrusives

Basic flows.—Microscopic examination of 40 thin sections reveals that many of the basic flows in the area are badly altered. The alteration consists essentially of albitization, uralitization, chloritization, and carbonation. Other investigations in the Upper Peninsula of Michigan (Butler and Burbank, 1929; Broderick, 1935; Cornwall, 1951) and across Lake Superior in Minnesota (Sandberg, 1938) suggest that alteration of Keweenawan flows in those regions was not so extreme as in the present area where less than half the diabasic basalts, only a very few of the melaphyres and porphyrites, and none of the amygdaloids remain unaltered.

In many cases the original character of the Wisconsin extrusives is very nearly obliterated by the metamorphic changes. However, relict textures and minerals permit interpretation of the primary rock. These relict phenomena include: (1) relict diabasic texture *i.e.*, tiny laths of albite are imbedded in or cut across amphibole or chlorite masses; (2) pyroxene remnants in otherwise completely chloritized and albitized rocks; and (3) relict phenocrysts of basic feldspar in albitized rocks. In addition, the abundant epidote granules in the now albitic feldspar suggest that the feldspar was formerly more basic. Alteration from the relatively fresh basic extrusives to the completely altered ones can be readily traced. The following alteration series, which was noted for one partially altered basic extrusive, can be duplicated many times for other samples from the area:



The two pyroxenes in this series are similar to those found in the associated gabbro.

Acidic flows.—The felsites and rhyolites are purplish to reddish in the hand specimen and exhibit a dense, aphanitic to very fine-grained texture with or without scattered pink feldspar phenocrysts. Rare micropegmatitic or granophyric interstices were noted under the microscope. Although potash feldspar occurs, the acidic feldspar is predominantly albite; hence, the felsites and rhyolites probably represent soda-rich flows. Minor amounts of hornblende magnetite, chlorite, sphene, apatite, and zircon occur in the groundmass. All feldspars are clouded by dustlike particles of kaolinite (?) and hematite; the latter gives the reddish color to the rock. This clouding effect is identical to that found in the granophyres.

For further details regarding the mineralogy, texture, and field relations of the basic and acidic flows and the following section on Keweenawan clastics, the reader is referred to an unpublished dissertation by the author entitled "Petrogenesis of a Gabbro-Granophyre Complex", written in 1951 and located at the University of Chicago.

Keweenawan Clastics

It is problematical whether the Keweenawan fragmental rocks of the area represent volcanic conglomerates, as defined by Wentworth and Williams (1932, p. 50), or true agglomerates. A mask of metamorphism has been imposed upon these rocks and has made interpretation of the primary rock difficult.

Ordinary sedimentary structures were not observed. Except for the rounded cobble- to boulder-size fragments, the dense reddish Keweenawan clastics do not have a sedimentary texture. Evidence of sand grains in the crystalloblastic matrix is lacking. Conceivably, the sand grains (if originally present) could have been destroyed by recrystallization as were the quartz grains in the Palms quartzite. Indeed, patches of epidote and twinned calcite have selectively replaced the matrix but have not destroyed the outlines of the rock fragments. Final decision regarding the origin of these coarse clastics should await further study.

Gabbro

General description.—Essentially, the gabbros are composed of only three or four minerals—plagioclase, pyroxene, magnetite, and at places, olivine. Weathered surfaces usually exhibit a rough, hackly surface with magnetite and pyroxene crystals projecting above the lath-shaped plagioclase hollows. With weathering, the plagioclases become white in contrast to the dark, dull magnetite and black to greenish gray pyroxene. In the hand specimen, most gabbros in the area are greenish gray and show a medium- to coarse-grained, hypidiomorphic granular, diabasic texture.

The relative proportion of minerals is variable. Everywhere, plagioclase is the most common mineral, except in pyroxene-rich bands in the banded gabbros. The feldspar is particularly predominant in the southern gabbro belt. Next in abundance is pyroxene, commonly of two types—augite and hypersthene. The ratios of the two minerals, plagioclase to pyroxene, ranges from 5:1 to nearly 1:1. Olivine ranges from zero to slightly over 20 per cent. Magnetite, present in every specimen, amounts to 4–5 per cent in most sections but is as high as nine per cent in one. In fresh specimens the remaining minerals form less than one per cent.

Mineralogy.—

Plagioclase occurs in subhedral crystals with an average composition of labradorite. Some specimens show zoning from bytownite to andesine or even more alkalic feldspar. (Anorthite content was determined by the Rittman zonal method and checked, in several cases, by the Federov method. A graph by G. C. Kennedy (1947, p. 562) was used to obtain the mol per cent of anorthite.)

Alteration of feldspar has produced sericite, epidote, clinozoisite, calcite, chlorite, and zoisite in decreasing abundance. In more altered specimens, chlorite or amphibole fill cleavage cracks and fractures in the feldspar. Specimens from the southern belt of gabbros show the most alteration; very few really fresh specimens were collected in this belt.

Hypersthene and augite commonly occur together as interstitial crystals molded around plagioclase laths and, in a few places, around anhedral olivine crystals. These pyroxenes may occur as independent crystals or as intergrowths with each other. Where intergrown, the host is generally hypersthene. The blebs, patches, and lamellae of clinopyroxene in the orthorhombic pyroxene have a common orientation resembling the "lamellar" or "graphic intergrowths" discussed by Hess and Phillips (1938), Walker (1940), Walker and Poldervaart (1949) and Poldervaart (1947a).

The dominant pyroxene is augite, as it is in the gabbro at Duluth (Grout, 1918d, p. 628). It is practically colorless and nonpleochroic. In addition to the customary prismatic cleavage, parting parallel (001) and (100) is common. Twinning is also present in places. A network of dark needles and tiny plates commonly indicates schiller structure. These needles and plates are apparently composed of sphene or magnetite or both. In one crystal, one set is oriented parallel to (100). In another, they are oriented parallel to the α-γ plane or (010).

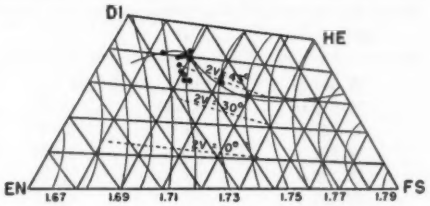


FIGURE 5.—PLOT OF CLINOPYROXENE COMPOSITIONS
Diagram after Poldervaart (1947b).

Approximate composition of clinopyroxenes was estimated from Figure 5 in which are plotted the measured values of 2V and γ.[1] The use of this diagram, taken from Poldervaart (1947b, p. 161) depends upon the assumption that these pyroxenes belong to the $CaSiO_3$-$MgSiO_3$-$FeSiO_3$ series of solid solutions and, thus, that the amounts of Al_2O_3, P_2O_5, TiO_2, MnO, and other impurities will be sufficiently low so as not to affect the optical properties in this system. Figure 5 reveals a clustering of points. In contrast to the Skaergaard intrusion (Wager and Deer, 1939, p. 241) and the Karroo dolerites (Walker and Poldervaart, 1949, p. 642–643), the clinopyroxenes in this investigation show only a small variation. This indicates that fractional crystallization has not been so effective as it has been in the Skaergaard and Karroo rocks.

The course of crystallization of pyroxenes as suggested by Hess (Poldervaart, 1947b) is also indicated in Figure 5. Since 2V has given unreliable results in the past, Poldervaart (1947b) emphasizes the use of γ-index curves and their intersection with the course of crystallization in order to obtain the most accurate results. Compositions of the clinopyroxenes determined by this method appear in Table 1, column 1; compositions represented by the points plotted in Figure 5 are shown in column 2 of the same table. Extinction angles

[1] The γ-index was obtained by adding .004 to γ' determined on (110) cleavage flakes—an approximation suggested by Walker and Poldervaart (1949, p. 630). The error involved in determining the indices should not be greater than $\pm$.005. Optic axial angles were determined directly on a universal stage and corrected for differences in refractive indices of the crystals and the hemispheres.

($Z \wedge c$) were determined wherever possible, and most are around 40°. Hess (1949, p. 630), however, points out that extinction angles have little determinative value in the clinopyroxenes of this solid solution series.

Hypersthene has nearly the same habit as the clinopyroxene and, like it, is also practically colorless; however, a faint pinkish pleochroism characterizes the orthopyroxene. The age relationships of the two minerals are undetermined. Some sections show only augite, and some show hypersthene rimming augite. These facts seem to indicate that hypersthene appeared after augite had begun crystallizing. This agrees with the observations of Hess (1941) who found that in diabases and dolerites augite normally begins to crystallize first.

Compositions of orthopyroxene crystals were determined from γ-index measurements, and checked by 2V determinations. A graph by G. C. Kennedy (1947, p. 564) was used to obtain the proportions of enstatite to ferrosilite. A later graph by Poldervaart (1950, p. 1076) was not used, since γ values and optic axial angles seemed to agree better with the graph by Kennedy. Several of the 2V measurements for intermediate orthopyroxenes were less than 51°, a value regarded by Poldervaart (1950, p. 1075) as the generally accepted minimum optic axial angle for these pyroxenes. However, he reports some orthopyroxenes with a 2V as low as 45°. Data on orthopyroxenes appear in Table 2. A range in composition from $En_{72}Fs_{28}$ to $En_{54}Fs_{46}$ was found. Alteration of both hypersthene and the clinopyroxene has resulted in the formation of uralite, biotite, and chlorite.

Olivine is present in some rocks and occurs in colorless anhedral crystals either imbedded in pyroxene or as independent crystals between laths of plagioclase. Serpentine is a common alteration product, particularly along fractures, and is accompanied by small magnetite aggregates. Several olivine crystals are rimmed by either green or reddish-brown biotite. Olivine compositions range from $Fo_{73}Fa_{27}$ to $Fo_{48}Fa_{52}$. Compositions were determined from β-index measurements which were then correlated with mol per cent fayalite in a graph presented by G. C. Kennedy (1947, p. 567). Optical data on olivines appear in Table 2.

Of the accessories magnetite is the most abundant. It occurs in euhedral to anhedral crystals both as interstitial material and as inclusions in every major rock-forming mineral. Skeleton crystals are present. Evidently the magnetite is actually magnetite-ilmenite since leucoxene alteration accompanies it. Another common accessory is apatite which is found as subhedral crystals in pyroxene and plagioclase. Brown biotite, generally associated with magnetite, is apparently primary in some sections. Pleochroic halos in some brown biotite indicate the presence of zircon. In addition, zircon euhedrons are imbedded here and there in the feldspar. Both brown and green hornblende occur in a few specimens. The brown amphibole appears primary, but the green seems to be an alteration product. Pyrite is a rare accessory, but when present it occurs in thin seams or bands of dispersed specks which cut directly across the rock.

TABLE 1.—COMPOSITION OF CLINOPYROXENES

Specimen No.	*2V*	γ	*Column 1*	*Column 2*
W-687-5	44°	1.731	$Wo_{32}En_{36}Fs_{32}$	$Wo_{32}En_{36}Fs_{31}$
W-239-0	42–51	1.730	$Wo_{32}En_{36}Fs_{31}$	$Wo_{33}En_{35}Fs_{30}$
W-267-1	39	1.721	$Wo_{39}En_{40}Fs_{21}$	$Wo_{32}En_{44}Fs_{23}$
W-187	52	1.720	$Wo_{40}En_{40}Fs_{20}$	$Wo_{40}En_{40}Fs_{20}$
W-320-0	49	1.720	$Wo_{40}En_{40}Fs_{20}$	$Wo_{39}En_{41}Fs_{20}$
W-397-1	54	1.719	$Wo_{40}En_{41}Fs_{19}$	$Wo_{42}En_{39}Fs_{19}$
W-423	50.5	1.718	$Wo_{41}En_{41}Fs_{18}$	$Wo_{40}En_{42}Fs_{18}$
W-427	38	1.718	$Wo_{40}En_{42}Fs_{18}$	$Wo_{33}En_{45}Fs_{22}$
W-130	41	1.717	$Wo_{41}En_{42}Fs_{17}$	$Wo_{35}En_{45}Fs_{20}$
W-646-1	50	1.717	$Wo_{41}En_{42}Fs_{17}$	$Wo_{40}En_{42}Fs_{18}$
W-136-5	42	1.716	$Wo_{41}En_{42}Fs_{17}$	$Wo_{36}En_{45}Fs_{19}$
W-461	45	1.716	$Wo_{41}En_{42}Fs_{17}$	$Wo_{38}En_{44}Fs_{18}$
W-842	49.5	1.715	$Wo_{41}En_{43}Fs_{16}$	$Wo_{40}En_{43}Fs_{17}$
W-264	48	1.708	$Wo_{41}En_{46}Fs_{13}$	$Wo_{40}En_{47}Fs_{13}$

Discussion of plagioclase, pyroxene, and olivine compositions.—Table 2 summarizes pertinent optical data and resulting compositions of plagioclase, orthorhombic pyroxene, monoclinic pyroxenes, and olivines found in the gabbro. If the results in Table 2 are expressed graphically, certain trends are indicated. For example, a plot of the ferrosilite content of orthopyroxene versus the anorthite content of the plagioclase gives a trend which suggests simultaneous iron and soda enrichment in the orthopyroxene and the plagioclase, respectively. A similar enrichment would be shown by a plot of olivine and plagioclase compositions. The trends would not be sharply

TABLE 2.—PYROXENES, OLIVINES, AND PLAGIOCLASES IN GABBROS

Number	Plagioclase		Monoclinic Pyroxene				Orthopyroxene				Olivine			
	Anorthite Content (%)		ext. ∠	2V	γ	Composition	sign	2V	γ	Composition	sign	2V	β	Composition
	Average	Range												
W-11	65	.....	?	?	1.723	$Wo_{27}En_{33}Fs_{24}$	....	...	....	.....	....	.....	....	.....
W-130	74	.....	41°	41°	1.717	$Wo_{41}En_{42}Fs_{17}$	(−)	?	1.703	$En_{71}Fs_{29}$	(−)	81° (av)	1.716	$Fa_{31}Fo_{69}$
W-136-5	.....	.....	42	42	1.716	$Wo_{41}En_{42}Fs_{17}$	....	...	....	.....	....	olivine alteration		
W-187	59	61–57	44	52	1.720	$Wo_{40}En_{40}Fs_{20}$	(−)	60°	1.706	$En_{69}Fs_{31}$	(−)	?	1.728	$Fa_{35.5}Fo_{64.5}$
W-239-0	71	75–66	40	42–51	1.730	$Wo_{32}En_{38}Fs_{31}$	....	...	....	.....	....	.....	....	.....
W-264	72	82–62	ca. 40	48	1.708	$Wo_{41}En_{46}Fs_{13}$	....	...	....	.....	(−)	85–90	?	.....
W-267-0	71	80–62	?	?	1.719	$Wo_{40}En_{41}Fs_{19}$	(−)	60	1.708	$En_{67}Fs_{33}$	(−)	84.5	1.719	$Fa_{32}Fo_{68}$
W-267-1	66	81–54	38	39	1.721	$Wo_{39}En_{40}Fs_{21}$	(−)	63	1.706	$En_{69}Fs_{31}$	(−)	83	1.723	$Fa_{34}Fo_{66}$
W-320-0	ca. 60	65–38	40	49	1.720	$Wo_{40}En_{40}Fs_{20}$	(−)	55.5	1.711	$En_{65.5}Fs_{34.5}$	(−)	?	1.749	$Fa_{46}Fo_{54}$
W-397-1	ca. 50	52–37	?	54	1.719	$Wo_{40}En_{41}Fs_{19}$	(−)	46	1.726	$En_{54}Fs_{46}$	....	olivine alteration		
W-423	52	58-albite	?	50.5	1.718	$Wo_{41}En_{41}Fs_{18}$	(−)	47.5	1.720	$En_{58}Fs_{42}$	....	.....	....	.....
W-427	65	75–48	41.5	38	1.718	$Wo_{40}En_{42}Fs_{18}$	(−)	48	1.717	$En_{61}Fs_{39}$	(−)	76	1.763	$Fa_{52}Fo_{48}$
W-461	58	Sl. zoning	44	45	1.716	$Wo_{41}En_{42}Fs_{17}$	(−)	$\gamma < 1.715$ present but rare		.....	(−)	75	1.740	$Fa_{42}Fo_{58}$
W-646-1	61	64–57	ca. 43	50	1.717	$Wo_{41}En_{42}Fs_{17}$	....	...	....	.....	....	.....	....	.....
W-687-5	54	62–10	?	61–40.5	1.731	$Wo_{32}En_{38}Fs_{30}$	....	...	....	.....	....	.....	....	.....
W-842	73	80–64	43	49.5	1.715	$Wo_{41}En_{43}Fs_{16}$	(−)	61	1.701	$En_{72}Fs_{28}$	(−)	85.5	1.709	$Fa_{27}Fo_{73}$
W1-181	61	64–57	?	?	1.718	$Wo_{41}En_{41}Fs_{18}$	(−)	present but rare		.....	present but altered and rare			

defined because of (1) errors in measurement of mineral composition, and (2) failure of the minerals to reach complete equilibrium.

The indication of simultaneous iron and soda enrichment agrees with laboratory investigations of silicate melts (Bowen, Schairer, and Willems, 1930; Bowen and Schairer, 1935) and with theoretical deductions derived from field observations (Fenner, 1929, p. 227; 1931; Buddington, 1943, p. 125). The similarity between silicate melt experiments and the present situation where simultaneous enrichment of iron and soda occurs suggests that the high magnesian and calcic members of the mafic and of the plagioclase series represent the higher-temperature members.

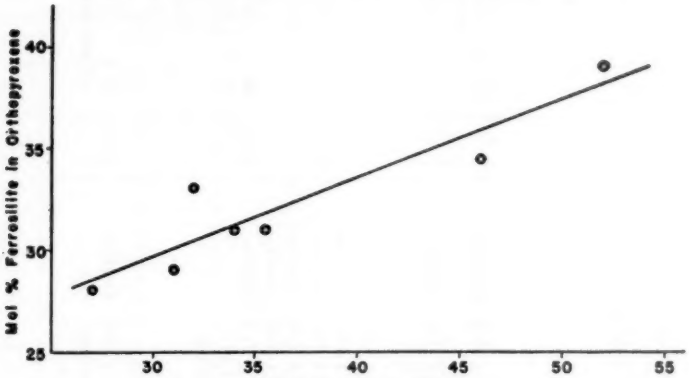


FIGURE 6.—RELATION OF COEXISTING ORTHOPYROXENES AND OLIVINES

Equation of line calculated by method of least squares is $Y = .39x + 18.0$. Coefficient of correlation is .931.

Other gabbros show similar results—*e.g.* the Skaergaard intrusion (Wager and Deer, 1939, p. 101), the Karroo dolerites (Walker and Poldervaart, 1949, p. 650–654), the Stillwater Complex (Hess, 1936, p. 199; 1939, p. 430–431), the dolerites of Tasmania (Edwards, 1942, p. 581 and 603), and the Palisade sill (Walker, 1940, p. 1069–1072). In all these, the writers indicate that iron enrichment accompanied by soda enrichment has resulted from fractional crystallization.

Whether or not the enrichment will yield ultimately to one of iron or of soda-rich differentiates is a matter of debate. This argument has been concisely summarized by Wager and Deer (1939, p. 293–297) and by Walker and Poldervaart (1949, p. 656–662). Bowen *et al.* (1930, p. 453) aptly state the problem thus:

> The crystallization theory requires the progressive concentration noted within each of these two crystallization series (feldspathic series and ferromagnesium series) taken individually, but it specifies nothing regarding the concentration of the two series relative to each other. In other words, theory has nothing to say as to whether the net result will be an increase in the absolute concentration in residual liquors of the iron-rich molecules of the one series or the alkalic-siliceous molecules of the other series. Indeed, the one or the other is bound to dominate. Which is it?

Figure 6 illustrates a remarkable correlation between fayalite enrichment in olivines and ferrosilite enrichment in orthopyroxenes. Inasmuch as the iron and soda enrichment of the pyroxene-olivine and plagioclase series reflects a temperature control, it is believed that the simultaneous enrichment of fayalite and ferrosilite shown by Figure 6 reflects the same control. In addition, the rather good correlation between fayalite and ferrosilite content suggests an approach toward equilibrium between these two minerals. This was unexpected, since in a basic magma it is generally presumed that olivine crystallizes prior to orthopyroxene, and, therefore, those crystals forming later should not necessarily be in equilibrium with crystals forming earlier.

Hence, an explanation is needed for the ap-

parent approach to equilibrium between olivine and orthopyroxene. Two possibilities seem to exist: (1) olivine and pyroxene crystallized simultaneously and, thus were in equilibrium with the melt and with each other; or (2) after crystallization of the two minerals there was a readjustment of iron and magnesium ions (a diffusion process) between them.

With reference to (1), Bowen and Schairer (1935, p. 187, 188, 211) find that, in the system MgO-FeO-SiO_2, there is a field in which olivine and pyroxene[2] may crystallize simultaneously with no resorption of olivine. This relation exists when more than 20 per cent of $FeSiO_3$ is present in the $MgSiO_3$-$FeSiO_3$ pyroxene series. A comparison of $FeSiO_3$ content of the orthopyroxene in the present investigation shows that this pyroxene may have formed in the field of simultaneous olivine and pyroxene crystallization.

With this assurance, the rocks containing olivine and orthopyroxene were again examined to see if textural relations supported the hypothesis of simultaneous crystallization. Evidences both for and against this hypothesis were recorded. Against it is the fact that quite a few olivines are rimmed by pyroxene indicating differing stages of crystallization. The existence of discrete crystals of olivine and pyroxene supports the hypothesis, however. Thus, although olivine obviously preceded pyroxene in some cases, textural relations suggest simultaneous crystallization in others.

A recent article by Ramberg and DeVore (1951) has a bearing on possibility (2). They consider an interaction between solid olivine and solid orthopyroxene as follows:

$$\underset{\text{forsterite}}{Mg_2SiO_4} + \underset{\text{ferrosilite}}{2FeSiO_3} \rightleftarrows \underset{\text{fayalite}}{Fe_2SiO_4} + \underset{\text{enstatite}}{2MgSiO_3}$$

The distribution of the iron and magnesium is believed to be in part the result of temperature control (p. 200) and in part due to non-ideality of the mixed crystals (p. 199). In the present case the gabbro represents a product of magmatic cooling (*see* section on Geological Interpretation) and, since there is a definite analogy between simultaneous iron and soda enrichment in the cooling of laboratory melts and in the gabbro, it is felt that temperature exerted the dominant control over the distribution of iron and magnesium. Thus, if diffusion between olivine and pyroxene occurred, it must have commenced shortly after their development in response to the still high temperatures of the partially solidified rock.

According to Bowen and Schairer (1935, p. 188, 213), olivine at equilibrium should be richer in iron than is pyroxene. This is true for the major part of Figure 6; however, with decreasing mol per cent of the iron silicate, there is a convergence in the difference between the iron content of olivine and pyroxene until at about 30 mol per cent they are equal, and even exhibit a reverse relationship. These observations are in complete agreement with those of Ramberg and DeVore (1951, p. 196) who find a reversal at about 35 mol per cent. Figure 6 represents an approximation of equilibrium conditions and since a reversal of olivines richer in iron to pyroxenes richer in iron is not a feature of the crystallization theory of Bowen, it is postulated that the present iron-magnesium ratios in the minerals are a result of post-crystallization readjustment. This redistribution was probably a consequence of decreasing temperature immediately after crystallization.

In summary, both simultaneous crystallization of pyroxene and olivine and post-crystallization readjustment are believed to have played a part in the present distribution of iron and magnesium in these minerals. As a result of fractional crystallization, different parts of the gabbro show greater iron to magnesium ratios than other parts. Those gabbros showing the greatest iron to magnesium ratios were formed at the lowest temperatures, analogous to laboratory melts. Each of these magma products with their varying iron-magnesium ratios was then further modified, by diffusion of iron and magnesium ions between olivine and pyroxene during cooling, in an attempt to reach equilibrium.

[2] The pyroxene obtained by Bowen was always clinopyroxene. However, as Bowen (1935, p. 200–201) pointed out, this pyroxene inverts to an orthorhombic form upon cooling or may even form as a primary precipitate if the temperature of crystallization is sufficiently low. Thus, a comparison between Bowen's experimental work containing clinopyroxenes and the present study showing orthopyroxenes is deemed proper.

Granophyre

General description.—From the descriptions in the literature (Grout, 1918e; Schwartz and Sandberg, 1940) there appears to be a close correspondence between the granophyres (red rocks) of Minnesota and the present area. Indeed, a description of the Wisconsin or Minnesota red rocks can be duplicated many-fold the world over. Particularly noteworthy, to name but a few, are the descriptions by Krokström (1932) for the granophyre associated with the Breven dolerite dike, by Wells and Wooldridge (1931) for the granophyres and red granites of Jersey, and by Strauss and Truter (1944) with regard to the Bushveld granites. The mafic content of the Minnesota red rocks is greater than that of the Wisconsin red rocks.

In the present area the granophyre is a massive, reddish, fine- to medium-grained, holocrystalline rock which shows an equigranular or granitoid texture in the hand specimen.[3] It is difficult in most sections to determine the average grain size, since individual grains which may be optically continuous are physically discontinuous due to intergrowth with some other mineral. However, it was recognized in the field that the eastern part of the area shows a coarser grain size than do the central and western parts. Some porphyritic types, with tiny pink feldspar phenocrysts in a very fine-grained groundmass, occur in the western half of the map area.

In outcrop the granophyre weathers to brick red. Fresh specimens also preserve the red color. Under the microscope this color is found to be a result of reddish, dustlike particles of iron oxide (hematite?) in the feldspar. Small, dark-green to black mafic spots are sparsely scattered throughout the hand specimen.

Microscopically, the texture shows considerable variation. Some degree of implication texture involving quartz and alkali feldspar is present in every rock, but the degree and pattern of implication is quite variable. Several types of this implication (granophyric) texture were found. These are the cuneiform, myrmekitic, irregular, radiating-fringe, and feathery or spherulitic types (Pl. 2).

The cuneiform type (Pl. 2, fig. 1) shows a regular pattern of wedge-shaped intergrowths of quartz and alkali feldspar. This is the ordinary micropegmatitic texture of most authors. The myrmekitic type (Pl. 2, fig. 2) shows elongated blebs or wormlike patches of quartz in feldspar. A sharp distinction between the cuneiform and myrmekitic types of intergrowth cannot always be drawn. In the case of the albite granite near Sparta, Oregon, Gilluly (1933, p. 68) also recognized a gradation in these textural types.

Wherever the intergrowths become less common and less regular, the type shown in Figure 3 of Plate 2 occurs. This irregular type still possesses a degree of intergrowth as exhibited by the optically oriented patches of quartz penetrated in places by optically continuous feldspars. Another characteristic type shown (Pl. 2, fig. 4) is the radiating-fringe type which consists of a central core of albite from which an intergrowth of quartz and alkali feldspar radiates outward. Toward the outer part of

[3] For the sake of simplicity, igneous textural terms are applied to the granophyre rather than their metamorphic equivalents. However, these terms are used primarily for descriptive purposes and have no genetic significance.

PLATE 2.—PHOTOMICROGRAPHS—RED ROCK

FIGURE 1.—CUNEIFORM TYPE OF IMPLICATION TEXTURE

Crossed nicols. ×33

FIGURE 2.—MYRMEKITIC TYPE OF IMPLICATION TEXTURE

Crossed nicols. ×33

FIGURE 3.—IRREGULAR TYPE OF IMPLICATION TEXTURE

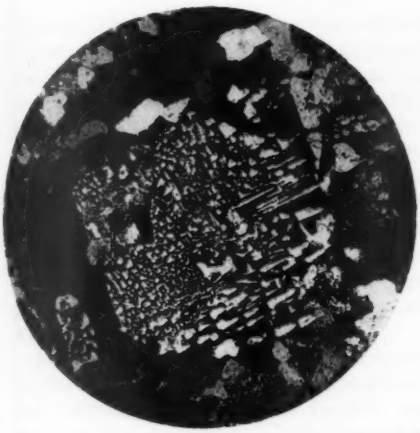
Crossed nicols. ×33

FIGURE 4.—RADIATING FRINGE TYPE OF IMPLICATION TEXTURE

Crossed nicols. ×33

FIGURE 5.—GRANOPHYRIC, SPHERULITIC, FELSITE PORPHYRY DIKE

Note feathery spherulitic intergrowth of quartz, and alkali feldspar radiating outward from resorbed turbid microperthite (left) and clear quartz crystals. Aphanitic groundmass. Crossed nicols. ×13



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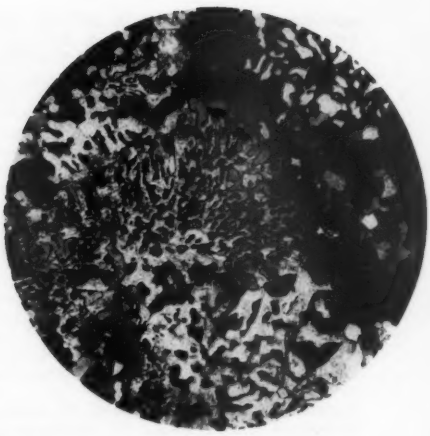

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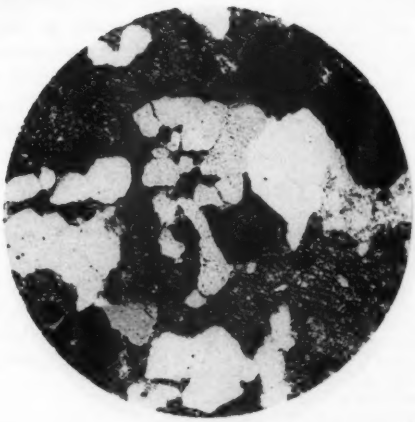

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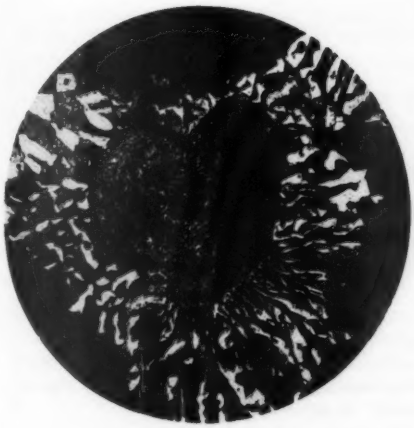

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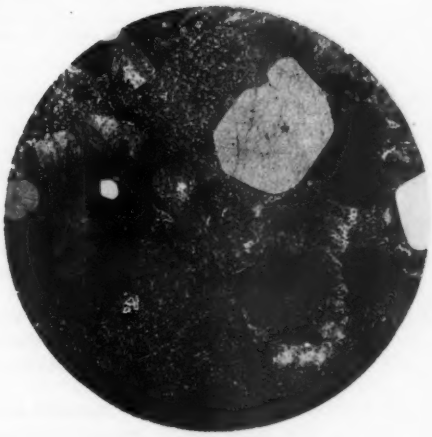
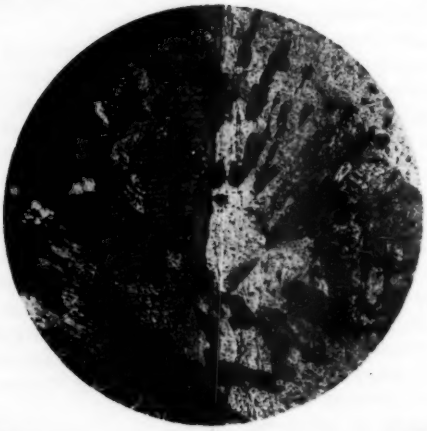


5

PHOTOMICROGRAPHS — RED ROCK

1

2



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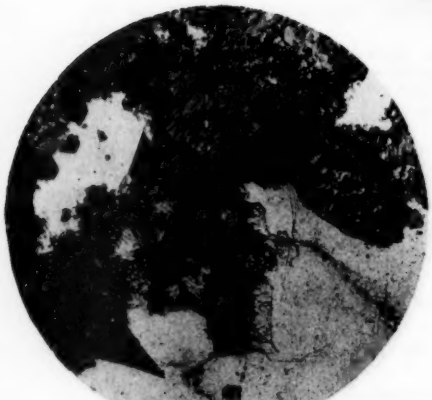

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PHOTOMICROGRAPHS — INTERMEDIATE ROCKS

this radial arrangement, the intergrowth may assume a cuneiform or myrmekitic character. On a much finer scale, the radiating-fringe type becomes the spherulitic or feathery type. This latter type exhibits an extremely fine intergrowth with individuals too small for optical measurements. One, two, or even three of the granophyric types of intergrowths may exist in one specimen. That they do not represent the same type of intergrowth cut at different orientations is shown by the fact that each may appear by itself in a thin section in which innumerable orientations occur.

After each thin section for the northern and southern red rock belts had been examined and described, the descriptions for each belt were arranged by locality in order from east to west. The irregular types of intergrowth occur exclusively in the eastern parts in both belts of granophyres. Toward the west the irregular types give way to more regular and persistent intergrowths of the cuneiform and myrmekitic types, and even the radiating and feathery or spherulitic types. Thus, there is a textural gradation of the implication texture along the strike of both belts. Such a gradation is not uncommon. For example, Walker and Poldervaart (1949, p. 669), Drescher-Kaden (1948, p. 172), and Rastall (1906, p. 270) report gradational relationships in the degree of development of granophyric texture.

Mineralogy.—

The granophyre is essentially a combination of quartz, potash feldspar, and albite. Other minerals amount to less than 10 per cent and generally less than 5 per cent. An abnormal deficiency of mafics was established by Rosiwal analyses of six specimens of granophyre. Five of these analyses are presented in Table 8. (The sixth is not listed because alkali determinations were not completed for the sample. The Rosiwal analysis of this sample, however, gave 37.1% quartz, 59.5% microperthite, 2.3% magnetite plus hematite, and 1.1% chlorite.)

Quartz occurs (1) as subhedral to euhedral phenocrysts, (2) as intergrowths with alkali feldspar, and (3) as anhedral interstitial granis. In every specimen (2) is present, but (1) and (3) may be absent. In contrast to the feldspars, the quartz is clear, with local undulatory extinction. Tiny inclusions of magnetite, acid feldspar, and iron-oxide dust are sparsely scattered in the quartz crystals. In many places an isolated feldspar fragment, in optical continuity with a bordering feldspar crystal, is found swimming in a sea of quartz.

Cloudiness of feldspar, a result of abundant dustlike inclusions of kaolinite and iron-oxide, hampers optical study. The dominant feldspar, however, is microperthite (potash feldspar patches appear to be slightly more abundant than the albite patches), and a small amount of plagioclase, either albite or oligoclase, is also present in some sections. At a few places the microperthite apparently grades to antiperthite. It is undetermined whether these crystals are really antiperthitic or are true microperthites cut in such a manner that the plane of the thin section is almost parallel to the albitic lamellae (thus exposing but small amounts of potash feldspar). However, in one thin section the feldspar seemed to be predominantly antiperthite regardless of crystal orientation.

Chemical data indicate that the ratio of potash feldspar to albite is nearly equal in most specimens. This holds true even for rocks in which microperthite was not identified with certainty. Some rocks show an intergrowth of alkali feldspars on a very fine scale, and, owing to clouding, the microperthitic character is difficult to verify.

A preferred orientation of albitic patches in microperthite was observed. On sections cut parallel to the (010) of orthoclase, the lamellae, blebs, patches, or thin, wedging films of albite are subparallel and form an angle of 62°–71° with the

PLATE 3.—PHOTOMICROGRAPHS—INTERMEDIATE ROCKS

FIGURE 1.—IRREGULAR EMBAYED TYPE OF ZONING

Shown by shading in large plagioclase crystal in bottom half of photograph. Crossed nicols. ×13

FIGURE 2.—ANASTOMOSING ACID FELDSPAR STRINGERS

Large plagioclase crystal in upper left quadrant. Large irregularly zoned crystal in Pl. 3, fig. 1 also shows stringer of acid feldspar cutting more basic plagioclase. Crossed nicols. ×13

FIGURE 3.—ORIENTED AMPHIBOLE RELICTS IN PLAGIOCLASE

Amphibole relicts (dark) in right half of photograph are imbedded in plagioclase (light gray). Crossed nicols. ×33

FIGURE 4.—QUARTZ REPLACEMENT OF FELDSPAR

Quartz (nearly white) in lower part of photograph has partially replaced twinned plagioclase crystal in lower central part of photograph. Former feldspar boundary indicated by row of dust inclusions in quartz to right of twinned feldspar remnant. Crossed nicols. ×33

FIGURE 5.—PRISMATIC APATITE NEEDLES

Note plagioclase cores rimmed by microperthite, and granophyric interstices. Crossed nicols. ×13

(001) cleavage. Manebach twinning of the microperthite produces a herringbone texture of the subparallel arrangement in several of these sections.

In sections paralleling (001), there is typically an irregular patchwork of albite and orthoclase (rarely microcline). A few crystals show patches and lamellae of albite oriented parallel to the (100) and/or the (010) face of the potash feldspar, or possibly arranged in en echelon fashion with the optic plane bisecting the obtuse angle formed by the en echelon pattern. On (001) surfaces, thin albite twin lamellae may be observed in the plagioclase patches. The potash feldspar in these surfaces exhibits a speckled or pitted effect believed to be the result of either extremely fine intergrowths of albite and orthoclase or submicroscopic twinning of potash feldspar.

In sections cut parallel to (100) of the microperthite, the patches of albite and orthoclase show rectangular shapes with sides parallel to the (001) and (010) cleavages. Generally, albite in these sections shows a checkerboardlike twinning of very fine albite and pericline twins.

The origin of perthites has been discussed by numerous investigators (Andersen, 1928; Barth, 1930, p. 137; Spencer, 1938; Goldich and Kinser, 1939; Higazy, 1949). In whatever manner the present perthites were formed—whether by simultaneous crystallization, exsolution, or replacement—the importance of crystal structure in microperthite development cannot be overemphasized. That structural control was dominant is exhibited by the relations of albite patches to definite crystal planes in the potash feldspar, and also by the close approach to identical crystallographic orientation of the intergrown albite and potash feldspar material. That these two intergrown minerals approach crystallographic and optical continuity is shown by: (1) the near parallelism of albite twinning in plagioclase and (010) cleavage in orthoclase; (2) the apparent continuity of (001) cleavage as it passes from one feldspar patch to the next; and (3) the agreement (within a few degrees) of the Z vibration directions in both feldspars. Clearly, if the potash feldspar is orthoclase, then some slight difference in crystallographic orientation must exist, because of the coexistence of a monoclinic and a triclinic structure.

The potash feldspar in the microperthite is commonly orthoclase, but many thin sections also show a grid twinning characteristic of microcline. Patches of orthoclase appear to grade into patches of true microcline. As a rule the patches of microcline do not possess the clouded effect so common in orthoclase. The inversion of orthoclase to microcline may be responsible for clearing of the dustlike particles of kaolinite and hematite.

Microcline grid twinning is particularly pronounced immediately adjacent to albite intergrowths. Away from these albitic patches and toward the center of the potash feldspar, grid twinning disappears. The occurrence of microcline immediately adjacent to albite lamellae indicates that the microcline development was promoted by either (1) exsolution of albite from orthoclase or (2) replacement of orthoclase by albite. In the case of exsolution, the orthoclase would have been metastable so that in or near areas of greatest albite accumulation *i.e.* in areas where migration of sodium and potassium ions produced the greatest disturbance, the ions of Al and Si in the disordered monoclinic orthoclase structure were excited sufficiently to overcome the potential barrier which had previously held the orthoclase metastable. Once the potential barrier was overcome, the ions rearranged themselves in their lowest free-energy state—the ordered triclinic modification, microcline. Where the least migration of ions occurred, the excitation of the Al-Si ions in orthoclase was insufficient to overcome the potential barrier. Replacement of orthoclase by albite would also produce the same effect *i.e.*, there would be greatest disturbance of the orthoclase lattice immediately adjacent to the albite replacements.

The presence of orthoclase rather than microcline at most places is indicated by 2V measurements. These may be somewhat variable within one thin section and even within one crystal. Optic axial-angle measurements vary as much as 8° (from 58° to 66°) in a single crystal. The 2V for orthoclase is generally slightly less than 70° but ranges from 55° to about 74°.

The plagioclase intergrown with potash feldspar shows a variable anorthite content.[4] Anorthite content of the plagioclase in microperthite decreases from east to west except for a few variations found in the far east end of the area (Fig. 7).

Some specimens of red rock exhibit cores or phenocrysts of albite or oligoclase. These cores or phenocrysts never amount to more than a few per cent of the rock. Where present as cores, the plagioclase is rimmed by microperthite. One rock showed zoned cores of plagioclase with an anorthite

[4] To determine anorthite content, extinction-angle values obtained either in sections perpendicular to the *a* axis or in sections which show maximum extinction perpendicular to (010) were plotted in the graph given by Kennedy (1947, p. 562).

content of 27 per cent in the central parts and 15 per cent in the outer parts. The cores show little of the turbidity which exists in the main mass of the rock and also possess wider twin lamellae than the albite in the microperthite. Alteration of the cores is generally to sericite and epidote. Sericitic alteration is common in the phenocrysts but may be absent in the groundmass. In one section, clear pools of orthoclase have replaced sericitized albite phenocrysts.

With the exception of magnetite and hematite ferromagnesian minerals are almost nonexistent in many red rock outcrops. Magnetite occurs in anhedral to euhedral interstitial crystals, as needles piercing the rock, as skeleton crystals associated with biotite and in places zircon, and as inclusions in every major rock-forming constituent. It is a common associate of chlorite and biotite. Oxidation along its borders and along cracks has produced hematite.

An iron oxide dust found in feldspars is believed to be hematite. This reddish dust is generally concentrated around the edges, along fractures, twin planes, and cleavages of alkali feldspars, and along the boundaries of plagioclase cores or phenocrysts and the groundmass. However, the greatest concentration is around magnetite crystals. Some microperthites show parallel rows of the reddish dustlike particles parallel to the *c* axis. The source of the iron oxide is a problem. Possibly at very high temperatures iron was able to enter the crystal lattice. Upon cooling it may have exsolved and formed hematite. Ramberg (1949, p. 30) reports absorption of about 0.5 per cent FeO and Fe_2O_3 in the feldspar lattice. Another possibility is that iron was introduced under hydrothermal or metamorphic conditions.

Both green and brown biotite are present. The green variety (nearly uniaxial) is more abundant. Biotite occurs in small flakes, shreds, or feltlike masses, and many are warped or bent, and have corroded boundaries. Spherulites of biotite occur in many specimens. Some yellowish cryptocrystalline patches occurring in feldspar may represent biotite remnants. Inclusions of biotite in feldspar are common.

Biotite in many places shows alteration to chlorite, and some flakes and shreds exhibit a fibrous alternating arrangement of the two minerals. Like the biotite, chlorite occurs in felty aggregates or spherulitic masses. Other ferromagnesian minerals include hornblende (most common), augite, and hypersthene. Augite with $2V = 57°$ and Z departs from *c* by 40° occurs in anhedral crystals in one specimen. In another specimen a trace of hypersthene was found.

Additional accessory minerals include apatite, pyrite, sphene, and zircon. Apatite is persistent and occurs as small inclusions in both feldspar and quartz. Pyrite is present in a few specimens, sparsely distributed in narrow bands which traverse the rock. Sphene is generally associated with the ferromagnesian minerals and may be an alteration product of them. Some sphene is found as inclusions in feldspar. Zircon occurs in subhedral to euhedral prismatic crystals, as inclusions in quartz or feldspar.

Peculiar mineral aggregates occur scattered throughout some specimens; these are generally composed of anhedral quartz-chlorite-magnetite

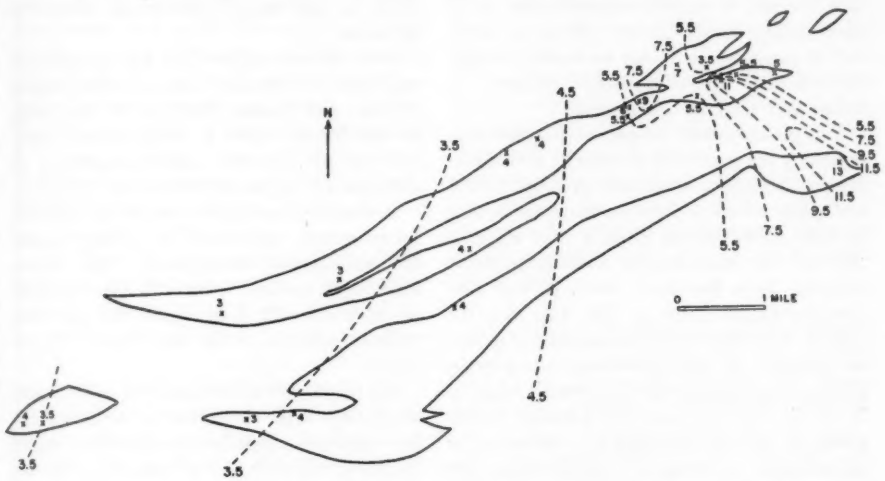

FIGURE 7.—CONTOUR MAP OF ANORTHITE CONTENT IN RED ROCK

Solid lines outline belts and patches of red rock. Dashed lines are contours of anorthite content. Contour interval 3.5, 4.5, 5.5, 7.5, 9.5, and 11.5 mol per cent anorthite. *X* indicates sample location. Adjacent to it is the anorthite content for that specimen. Values of anorthite content in each specimen obtained by averaging three or more extinction-angle measurements in at least three crystals.

aggregates which apparently represent alteration products of an earlier ferromagnesian mineral.

Origin of granophyric texture.—The origin of granophyric texture is a much discussed and still unsolved question. If primary, is this texture a result of eutectic crystallization or of simultaneous crystallization without a eutectic? If secondary, is it due to deuteric action, hydrothermal alteration, solid diffusion, or some other granitization process?

The present study reveals new information regarding the possibility of eutectic crystallization, a concept championed by Vogt (1930) and many others. Recent evidence shows that eutectic crystallization is not a valid explanation of the micrographic and granophyric textures. Both Backlund (1946, p. 112) and Drescher-Kaden (1948, p. 109) state that the finding of a wide variation in amounts of quartz to feldspar in quartz-feldspar intergrowths eliminates the possibility of a eutectic relation. Textural evidence given by Drescher-Kaden (1948, p. 139–147, 203–209) also indicates the unlikelihood of eutectic crystallization for many of these intergrowths.

In the present case, the occurrence of granophyric intergrowths of quartz with feldspars that show a range in anorthite composition (Fig. 7) precludes the possibility of eutectic crystallization. Thus, we are left with the choice between simultaneous crystallization without eutectic or some sort of secondary origin.

That a secondary origin for granophyric intergrowths is possible is shown by the presence of these textures in metamorphic rocks—*e.g.*, in meta-quartzites (Drescher-Kaden, 1948, p. 203–205). Even if the origin is secondary, many writers disagree on the processes involved. For example, Bowen (1922, p. 192) states that the graphic structure in many examples is probably a result of reaction between the phases during solidification. This is, in a sense, secondary. Colony (1923, p. 174) believed that micropegmatite was the result of deuteric (magmatic end-stage) phenomena. Drescher-Kaden (1948, p. 249) concludes that hydrothermal solutions were responsible for the quartz-feldspar intergrowths. Other writers indicate the potency of granitization processes to form such textures.

These intergrowths may also be formed through crystallization of a melt (Reynolds, 1937, p. 268, 275; Walker and Poldervaart, 1949). Also highly suggestive of a primary origin is the occurrence of these intergrowths in unaltered diabases. In addition, Johannsen (1932, p. 288) records granophyric texture in extrusives.

Some writers suggest that the intergrowths may have both primary and secondary origins (Strauss and Truter, 1944, p. 53). According to Wahlstrom (1939, p. 695), crystal forces can produce the same features regardless of replacement origin or simultaneous growth.

In the present example some indirect evidence favors quartz replacement of feldspar to give the granophyric intergrowth. Since quartz undeniably replaced feldspar in the intermediate rock zone (Pl. 3, fig. 4), it may also have replaced feldspar in the main mass of granophyre.

Another aspect to be considered in the granophyre belts is the correlation or lack of correlation between plagioclase composition, degree of graphic intergrowth, and grain size. Although there is an over-all decrease in grain size from east to west, closer examination reveals no real correlation, from one specimen to the next, between grain size and degree of intergrowth or between grain size and composition. There is a fair correlation between plagioclase composition and degree of intergrowth. This correlation suggests that the degree of intergrowth is influenced by the feldspar lattice. With decreasing anorthite content in the alkalic feldspar the lattice may change in such a manner as to cause a relatively low surface tension between the quartz and alkalic feldspar, and thus promote intergrowth. Whether this influence of structure prevailed during primary crystallization (simultaneous crystallization) or during replacement cannot be ascertained. If formed from a melt, the granophyric intergrowths may also depend upon rate of cooling or pressure (Rastall, 1906, p. 271).

Dikes.—One particular red rock dike warrants special attention. This dike occurs in the north of sec. 30, T. 45 N., R. 4 W. and cuts through felsite flows. In the hand specimen, the rock shows medium-grained quartz and pink feldspar phenocrysts embedded in a reddish aphanitic groundmass. The pheno-

crysts amount to approximately 35 per cent of the rock (Pl. 2, fig. 5). The phenocrysts of feldspar are microperthitic and resemble the feldspar crystals in the coarser red rock a short distance to the east. Both the quartz and the microperthitic phenocrysts show resorption phenomena, and spherulitic growths of acid feldspar occur around many of these porphyritic crystals. In places the radiating fibers of feldspar are intergrown with quartz, giving a feathery implication texture. The phenocrysts with their spherulitic overgrowths are set in an extremely fine-grained (less than .014 mm) groundmass which rarely exhibits minute quartz-feldspar intergrowths.

That this dike is genetically related to the main red rock body is shown by the identical appearance of the microperthite in both rocks, by the hematite clouding, and by the same relative amounts of minerals (particularly the deficiency of mafics). Regarding this relation of granophyre and granophyric spherulitic felsite porphyry, one is reminded of the description by Hall (1932, p. 244 and 247) of a transition "from felsite through spherulitic granophyric felsite, granophyric felsite, and fine-grained granophyre, to ordinary granophyre."

Intermediate Rocks

General description.—The narrow zone between the red rock and gabbro shows the greatest range in mineralogy and texture of any of the Keweenawan rocks in the belted complex. It ranges from fine- to coarse-grained, from xenomorphic to hypidiomorphic granular, from equigranular to inequigranular. It varies from predominantly red adjacent to the granophyre to predominantly dark gray adjacent to the gabbro.

Texture is also characteristic of a certain position in the intermediate zone. Away from the red rock, the texture varies from that simulating the granophyre to that duplicating the gabbro, *i.e.*, from xenomorphic granular with scattered small prisms of amphibole through hypidiomorphic granular with coarse-bladed amphibole crystals to a pseudo-diabasic rock with plagioclase laths cutting amphibole boundaries. The pseudodiabasic texture passes gradually into the regular diabasic texture of gabbro. In this sequence, micropegmatite decreases from the dominant material to rare interstitial patches. The grain size changes from that of the granophyre (about 1 mm) to that of the gabbro (about 3 mm). In the northern belt these variations are found on both sides of the red rock tongues which project into the gabbro (Pl. 1).

In one sequence of thin sections taken across the "transition" zone, a cataclastic texture was exhibited, but it is not present in most intermediate rock specimens.

Mineralogy.—A typical mineralogical sequence from red rock to gabbro is shown in Table 3. This table shows the overall character of the "transition" zone and has therefore been generalized.

The feldspars illustrate the variability of minerals across the intermediate rock zone. Three distinct occurrences were identified: (1) cores of plagioclase rimmed by microperthite or albite, (2) discrete crystals of plagioclase or microperthite, and (3) intergrowths (microperthite).

The occurrence of plagioclase cores is striking. As Bailey and Thomas (1924, p. 329) pointed out with regard to examples on Mull, they appear out of equilibrium with their surroundings. Near the gabbro the central parts of these cores approach labradorite in composition. Toward the granophyre, they become richer in soda. The cores become increasingly abundant and larger toward the gabbro. Varying degrees of alteration coincide with changes in composition. For example, if the central parts of the cores are andesine, they do not show such great alteration as the outer oligoclase or albite parts of the cores. The central parts may remain clear or show only beginning stages of sericite alteration. With a decrease outward in anorthite content, sericite alteration increases and is accompanied by epidote. Toward the outer parts, the more albitic portions may show kaolinite and some hematite dust.

The zoning of these cores is not of the ordinary type that shows various stages of regular concentric growth. Instead the zoning appears irregular and embayed (Pl. 3, fig. 1). In places, these embayments seem to favor certain twin planes over others.

In addition, anastomosing stringers of clear, more acidic feldspar cut many of these plagioclase cores (Pl. 3, fig. 2). In a few cases where the stringers are so numerous, basic feldspar remains as isolated patches. Twinning in these cores is partially obliterated by the stringers and by the irregular type of zoning.

The rims show little of the sericite and epidote alteration but are clouded by kaolinite and hematite dust. These rims are microperthitic close to the red rock, but are albitic or antiperthitic near the gabbro. One specimen revealed sericitic ghosts after plagioclase imbedded in an aggregate of microperthite.

Much of the feldspar shows corrosion by quartz.

TABLE 3.—MINERALOGICAL SEQUENCE FROM RED ROCK TO GABBRO

	Red Rock	*Intermediate zone*		*Gabbro*
		Nearer red rock	*Nearer gabbro*	
Plagioclase as cores	Albite (An_4) ↓ kaolinite sericite	Oligoclase-albite (An_{30}-An_4) ↓ epidote kaolinite sericite	Andesine An_{45}-albite ↓ chlorite epidote sericite	None
Plagioclase as distinct crystals	None	None	Albite-oligoclase (An_{30}-albite) ↓ epidote kaolinite sericite	Labradorite-andesine (An_{65}-An_{45})
Feldspar intergrowths	Microperthite (orthoclase and albite (An_4)) ↓ kaolinite	Microperthite (orthoclase and albite (An_4)) ↓ kaolinite	Microperthite (orthoclase and albite (An_4)) ↓ kaolinite	Rare microperthite
Quartz	Quartz-feldspar intergrowths abundant	Quartz-feldspar intergrowths common	Quartz-feldspar intergrowths persistent	Quartz-feldspar intergrowths rare
Ferromagnesian minerals	Chlorite	Blue-green amphibole ↓ biotite, magnetite, yellow cryptocrystalline masses, chlorite	Blue-green (variable) amphibole ↓ biotite magnetite titanite	Pyroxene ↓ uralite (blue-green) magnetite
Common accessories	Hematite, magnetite, and zircon	Hematite, magnetite, zircon and apatite	Apatite, hematite, magnetite, and zircon	Apatite, hematite, magnetite, and zircon

That quartz actually has replaced feldspar is illustrated in Figure 4 of Plate 3. Feldspar, too, has replaced ferromagnesian minerals. Reference to Figure 8 and to figure 3 of Plate 3 shows oriented pyroxene and amphibole remnants, respectively, in plagioclase.

The ferromagnesian minerals show remarkable variations. The dominant mafic mineral is amphibole. It occurs in arborescent groups or in coarse-bladed crystals (up to 23 mm in length) which are warped in places. It also commonly occurs in fibrous patches or felty aggregates. Most crystals show a ragged outline and contain numerous small irregular magnetite inclusions.

Much of the amphibole is pseudomorphous after pyroxene. Pyroxene parting is still preserved in many amphibole crystals, and pyroxene remnants are common particularly near the gabbro. In addition, some amphibole crystals show the same type of "graphic" intergrowth exhibited by pyroxenes. These amphibole intergrowths appear as pseudomorphs of earlier intergrowths of orthorhombic and monoclinic pyroxenes.

The microscope reveals a very inhomogeneous hornblende even within one specimen and one crystal. It is extremely variable in pleochroism, 2V, and other optical properties. Although the general pleochroic formula for most amphiboles in the "transition" zone is X = greenish yellow, Y = olive green, Z = bluish green, there are

striking variations in the shades of these colors. Perhaps the most notable variability of colors is in the Z direction. Here the shades vary from greenish brown through green to bluish green, and even blue. An amphibole with a greenish core is commonly fringed by a blue-green zone. Toward the granophyre, the blue-green amphibole is more

FIGURE 8.—SKETCH OF PYROXENE RELICTS IN PLAGIOCLASE

Plagioclase (stippled) badly altered. Orientation of two pyroxene groups indicated by discontinuous lines which represent cleavage trace. ×20.

pronounced than elsewhere. As an example of the exceptional variation in optical properties, a single hornblende crystal showed deep blue rectangular patches intermingled with ordinary green amphibole patches. One blue patch showed a 2V of 33°, whereas an adjacent patch (same crystal) showed a 2V of 71°. In general, the deeper the blue the smaller the optic axial angle. This blue is probably due to a soda-rich hornblende (optical properties of hornblende are present). The diagram by Winchell (1945, p. 37, 41) shows that an amphibole with 2V = 30° would be rich in the $NaCa_2Fe_5AlSi_7O_{22}(OH)_2$ molecule.

Amphibole has altered to green biotite or fibrous blue-green chlorite. Epidote was found in a few amphibole crystals. In many places aggregates of quartz, chlorite, magnetite, and sphene may represent alteration products of amphibole. Yellowish cryptocrystalline patches (possibly remnants after hornblende or biotite) are imbedded in many plagioclase crystals.

Of the accessories, apatite shows the most remarkable occurrence. It is found in prismatic needles up to 10 mm cutting across all major mineral boundaries. Apatite crystals become particularly abundant in the intermediate rock adjacent to gabbro (Pl. 3, fig. 5). Magnetite, the most abundant accessory, is present as anhedral to euhedral crystals, as skeleton crystals, as acicular crystals, and as ragged inclusions in all chief rock-forming constituents. Other less common accessories include biotite, pyrite, sphene, and zircon. Rather surprisingly, anhedrons of blue-gray tourmaline were found in one sequence of specimens collected across the intermediate rock zone. This mineral was not found in either the gabbro or the red rock.

The "transition" zone can be matched in many gabbro-granophyre areas or for that matter, in many areas of gabbro and acidic rocks of any type. Bastin (1938) described the intermediate rock zone between the gabbro and red rock of famed Pigeon Point, Minnesota. Its characters are nearly identical to those in the present area. The description of the epidolerite of the Breven dike (Krokstrom, 1932) shows marked similarities to the intermediate zone. Adjacent to an alkaline rock mass, the Salem gabbro-diorite of Massachusetts grades into a rock containing quartz and alkaline feldspars with large bladed amphiboles (Clapp, 1921, p. 40) like those in the "transition" zone of the present area.

CHEMICAL DATA

Rock Analyses

Presentation of data.—Chemical analyses of five specimens taken across the "transition" zone (Table 4) are plotted in a graph (Fig. 9) showing the variation of oxide content across the zone. The specimens were collected from the large outcrop (Pl. 1) in the extreme southeast corner of sec. 26, T. 45 N., R. 5 W. Specimen W-687-2 represents the approximate position of the contact (contact covered). W-687-1 is a specimen of the red rock. The transition zone where the specimens were collected is about 85 feet wide. Table 5 lists the norms and modes of the rock samples whose analyses appear in Table 4.

Discussion of analyses.—Table 4 and Figure 9 reveal significant variations in TiO_2, FeO, MnO, and P_2O_5. All four of these oxides reach maximum values in specimen W-687-4. Note that this specimen belongs to the intermediate zone but lies closest to the gabbro. Such maxima were found by Collins (1934, p. 145, 148) for FeO and TiO_2 in the Sudbury nickel irruptive. Significantly, these peaks also occur near the gabbro-"transition" zone "contact." Collins does not show graphs for variation of MnO and P_2O_5.

In the Skaergaard intrusion, chemical analy-

ses of the layered rocks show maxima in iron, TiO_2, P_2O_5, and MnO—all near the same depth interval in the intrusive (Wager and Deer, 1939, p. 133, Pl. 27). Walker and Poldervaart (1949, p. 652–654) find that, in the Karroo dolerites, as the $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100$ ratio approaches 80, TiO_2 increases sharply. This corresponds closely with the present situation (Table 4). These authors also found that FeO reached a peak at a slightly lower ratio of about 65. Another remarkable sequence of rocks, those of the Greenstone flow in Michigan (Cornwall, 1951, p. 156–158) also show characteristic peaks. Again, these are of the oxides FeO, TiO_2, P_2O_5, and MnO, and again they occur in nearly the same rock specimens. A plot of oxide content versus per cent solidified (Cornwall, p. 157, 158) shows peaks for FeO and P_2O_5 for the same "percentage solidified" and peaks for TiO_2 and MnO for another "percentage solidified"; the difference in "percentage solidified" between these two pairs of peaks amounts to only 4 per cent.

The origin of the peaks in the Sudbury irruptive still remains a problem. In the three examples—Skaergaard intrusion, Karroo dolerites, and Greenstone flow—we are dealing with true igneous rocks; therefore, the maxima in those rocks are the result of magmatic processes.

For comparative purposes, a triangular diagram (Fig. 10) was constructed, with FeO, MgO, and $K_2O + Na_2O$ at the three corners. In it are plotted the five analyzed specimens. The trend of the variations from gabbro to granophyre is similar to those trends found in the Breven and Hällfors dikes and the Skaergaard liquid (both of these curves are shown in Wager and Deer, 1939, p. 314, 315). The trend is also similar to that shown by the Karroo dolerites and the Paleozoic quartz dolerites and tholeiites of Scotland and northern England (Walker and Poldervaart, 1949, p. 657, 659). In addition, Tilley (1950, p. 48) and Cornwall (1951, p. 167, 169) present triangular diagrams of the Duluth and Logan sills and of

TABLE 4.—CHEMICAL ANALYSES OF GRANOPHYRE, INTERMEDIATE ROCKS, AND GABBRO
Analyst: H. B. Wiik, Helsinki, Finland

	Red rock	*Intermediate rock*			*Gabbro*
	W-687-1	W-687-2	W-687-3	W-687-4	W-687-5
SiO_2	74.34%	70.20%	63.19%	50.11%	47.91%
TiO_2	0.32	0.56	1.41	3.43	3.00
Al_2O_3	10.01	12.33	12.22	12.73	15.46
Fe_2O_3	5.17	3.12	3.67	5.17	2.31
FeO	0.80	2.63	4.71	9.97	9.42
MnO	0.02	0.06	0.13	0.32	0.22
MgO	0.23	0.76	1.68	3.73	5.42
CaO	0.20	0.90	3.57	7.30	10.24
Na_2O	4.30	4.17	4.30	2.33	2.50
K_2O	3.20	3.57	2.26	1.76	1.09
P_2O_5	0.06	0.06	0.40	0.47	0.24
H_2O^+	0.98	1.15	1.96	2.35	1.89
H_2O^-	0.06	0.15	0.26	0.13	0.03
Total	99.69	99.66	99.76	99.80	99.73
Sp. Gr.	2.625	2.68	2.79	2.99	2.96
$\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100$	96.29	88.33	83.30	80.23	68.40
$\frac{K_2O + Na_2O}{CaO + K_2O + Na_2O} \times 100$	97.40	89.58	64.76	35.91	25.96

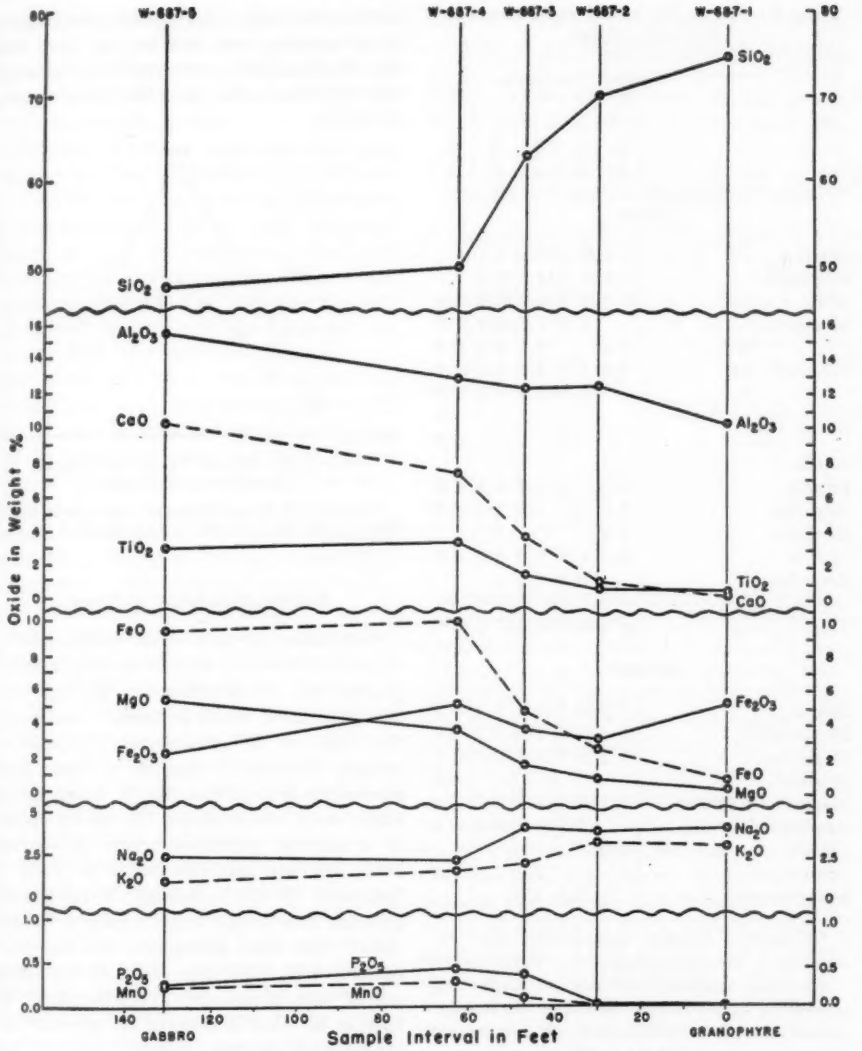

FIGURE 9.—OXIDE VARIATIONS ACROSS INTERMEDIATE ZONE

the Greenstone flow which further illustrate the same trend.

Thus, a comparison of the variations in the rock analyses in Table 4 with igneous areas suggests that magmatic differentiation was responsible for the gabbro-granophyre sequence. However, Reynolds (1946; 1947a) advances the hypothesis of metasomatism to explain such peaks or "geochemical culminations" (the term used by Reynolds for an increase of any constituent beyond the amount present in the parent rock or the adjacent granite). Reynolds (1946) cites seven areas in which granitic intrusions come in contact with earlier basic igneous rocks. In these areas the basic igneous rocks near the granite commonly show

TABLE 5.—NORMS AND MODES OF CHEMICALLY ANALYZED ROCKS

	Red rock	*Intermediate rocks*			*Gabbro*
	W-687-1	W-687-2	W-687-3	W-687-4	W-687-5
Norms					
Quartz	37.5	29.0	20.8	8.7	....
Orthoclase	18.9	21.1	13.3	10.6	6.7
Albite	33.5	35.1	36.2	19.9	21.0
Anorthite	....	3.6	7.5	18.9	27.8
Pyroxene* Wo	0.1	....	3.1	6.0	9.3
Pyroxene* En	0.6	1.9	4.2	9.3	12.2
Pyroxene* Fs	....	1.5	3.4	8.8	9.8
Olivine Fo	....	....	....	....	1.0
Olivine Fa	....	....	....	....	0.8
Acmite	2.3	....	....	....	....
Ilmenite	0.6	1.1	2.7	6.5	5.8
Magnetite	1.6	4.4	5.3	7.4	3.3
Hematite	3.4	....	....	....	....
Apatite	0.3	0.3	1.0	1.0	0.3
Corundum	....	0.3	....	....	....
Total	98.8	98.3	97.5	97.1	98.0
Modes†					
Quartz	34.9	26.5	14.4	9.1	3.6
Microperthite	58.2	51.5	7.0	4.2	0.1
Plagioclase	1.2	9.7	51.3	37.5	62.7
Pyroxene	....	....	....	....	4.5
Magnetite + Hematite	4.7	4.9	8.2	6.7	3.6
Amphibole	....	7.1	13.2	38.9	24.4
Chlorite	1.0	....	....	....	....
Apatite	....	....	2.2	3.6	1.1
Miscellaneous	....	0.3	3.7	....	....

* Normative pyroxene compositions were determined in the manner suggested by Barth (1931).

† Rosiwal analyses were run on single thin sections for each specimen. Hence, for the coarser-grained rocks, W-687-3 through W-687-5, the modal determinations are probably not highly accurate.

geochemical culminations of the cafemic constituents and one or more of the minor constituents TiO_2, P_2O_5, and MnO. The culmination of these constituents has been termed a "basic front" by Reynolds (1947a, p. 211); she states: "A front occurs wherever there is a diffusion limit marked by a change in the mineral assemblage."

Hence, for an explanation of the geochemical culminations in the intermediate zone between the granophyre and gabbro, we have two plausible hypotheses—one involving magmatic differentiation and the other metasomatic alteration.

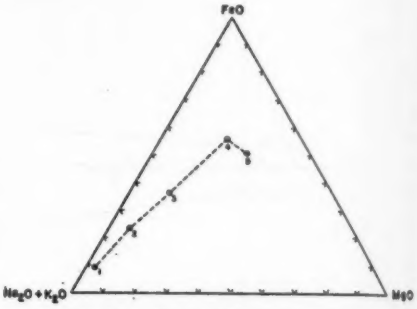

FIGURE 10.—OXIDE TREND IN GABBRO-GRANOPHYRE COMPLEX

Points 1, 2, 3, 4, and 5 represent analyses W-687-1, W-687-2, W-687-3, W-687-4, and W-687-5, respectively.

Isotopic Abundances of Oxygen

Presentation of data.—The distribution of oxygen isotopes in a suite of samples from the present area was determined in the hope that this information would be useful in indicating the origin of the granophyre. This suite of samples included a sequence of rocks from granophyre into gabbro. Dr. S. R. Silverman kindly made the analyses. His results appear in a recently published article (Silverman, 1952) and are also reproduced in Table 6. Specimens W-137-1 through W-137-4 were collected at intervals across a narrow "transitional" zone from granophyre into the intermediate rock. This zone occurs in the south-central part of sec. 25, T. 45 N., R. 5 W. Sample W-137-2 is believed to represent the approximate contact (contact covered) between granophyre on the south and the intermediate zone on the north. To the north, specimens W-137-3 and W-137-4 lie at intervals of 1 and 7 feet, respectively, from the contact rock. Sample W-136-4 lies 2 feet south of the contact. The two specimens, gabbro W-187 and granophyre W-136-4, represent unmodified end members in this sequence. The results are expressed as deviations in parts per thousand from a standard sample.

Table 6 shows a regular increase in the O^{18}/O^{16} ratio from gabbro through the intermediate zone into the red rock. In other words, O^{16} becomes relatively more abundant from the red rock toward the gabbro.

Discussion of isotopic variations.—One may well question the mechanism that produced the observed variations in isotopic abundances of oxygen. Silverman (1952), after effectively eliminating such mechanisms as the Soret effect, gravitational control of diffusion, and assimilation or melting of gabbro by a granophyric melt, believes that the variations are best explained by transfer or diffusion of material in the solid state. His conclusions are based largely upon chemical considerations; however, as is so often the case, chemical data may be interpreted in several ways. On the basis of chemical data alone, Silverman's explanation does not appear to be the only plausible explanation of the observed isotopic variations. If we neglect for the moment additional field and laboratory data, several mechanisms appear capable of explaining the observed variations: (1) metasomatism, (2) mechanical mixing, and (3) fractional crystallization.

Chemical kinetics tells us that the rate of migration of isotopes is inversely proportional to the square roots of their masses (Glasstone, 1946, p. 155; Thomson *et al.*, 1921, p. 88; Lindemann, 1921, p. 102, 103). Thus, the lighter O^{16} isotope should migrate faster than the heavier O^{17} isotope, and increase in abundance relative to the heavy isotope in the direction away from the source of migrating material. If the direction of migration was from granophyre to gabbro, the experimentally determined oxygen isotope ratios correspond with the results expected from theory. This is, indeed, the likely direction of migration if the granophyre is considered either as a center of granitizing action or as a melt promoting contact metasomatism. Thus, metasomatic replacement is a plausible explanation of the observed variation.

Mechanical mixing could originate in ways other than assimilation, which has been ruled out by Silverman. This mixing could occur through *lit-par-lit* injection, or fissure- or fracture-filling. If we consider an original red rock magma, high in O^{18}/O^{16} and a solid gabbro, lower in O^{18}/O^{16}, a partial mixing of the two in either of the suggested manners will give rise to a product with an intermediate value of O^{18}/O^{16}. This is exactly what we find in the intermediate zone; hence, partial mixing is also a possible explanation. (It should be remembered that partial mixing of two liquids would produce the same result.)

TABLE 6.—OXYGEN ISOTOPE ABUNDANCES IN THE BAYFIELD COUNTY GRANOPHYRE-GABBRO SEQUENCE

Sample and Number	O^{18}/O^{16} 0/100	*Average error* 0/100	*Number of determinations*	*cc O₂/mg. sample*
Gabbro, No. W-187	6.1	0.4	4	0.28
Intermediate rock, No. W-137-4	7.7	0.0	2	0.33
Intermediate rock, No. W-137-3	8.4	0.3	4	0.35
Contact rock, No. W-137-2	8.7	0.3	4	0.36
Granophyre, No. W-137-1	9.0	0.2	3	0.39
Granophyre, No. W-136-4	9.9	0.8	3	0.36

Lastly, let us consider the possibility that the red rock has originated through fractional crystallization and that the intermediate rock represents an intermediate step in the process. This, of course, suggests that the temperature of formation of the intermediate rock would be somewhat above 600°C. Although there is doubt whether isotopic fractionation would occur at this temperature, consideration of pressure and viscosity effects suggests that it would be possible for O^{18} to become preferentially enriched in the residual liquid rather than O^{16}.[5] For example, Bridgman (1950) has shown that, with extreme pressures, viscosity becomes so great that diffusion of material to crystallization nucleii may be prohibited during the period of cooling.[6] Thus, with somewhat

[5] In speaking of O^{18} and O^{16} in a melt, the writer means in combination with Si, since silicate melts are known to possess silicate ions rather than individual oxygen and silicon atoms.

[6] Bowen (1921) previously determined diffusion rates in silicate melts and found that diffusivity is inversely proportional to the viscosity. Also Vogt (1930, p. 180) states: "The greater the viscosity, the less the power of the molecules to . . . [orient] . . . themselves. . . ."

less pressure and less viscosity, there must be a stage at which only the more active ions will migrate toward crystallization centers. These active ions will then be taken into the crystal lattice while the less active ions remain behind, incapable of diffusion. Therefore, we have a situation in which the more active O^{16} ions would crystallize out leaving the O^{18} ions behind.

Availability of ions containing the O^{18} and O^{16} isotopes must also be considered. Thomson (*et al.*, 1921, p. 88), in discussing isotopes of chlorine in reactions between chlorine and arsenic and between HCl acid and an alkali, stated that "... a larger proportion of the faster molecules than of the heavier ones would strike against the absorbing substance and be absorbed, so that the gas left behind would gradually increase in density." In an analogous manner, perhaps a larger proportion of the faster migrating O^{16} ions than of the heavier O^{18} ions would contact the crystallization nucleii in a melt and be "absorbed," so that the liquid left behind would gradually increase in the heavier constituent.

Still another factor may control the O^{18}/O^{16} distribution during fractional crystallization. Perhaps the O^{18} to O^{16} ratios depend upon the lattice-energy requirements of the individual minerals. That is, in response to the energy requirements, a specific mineral may have an affinity for an O^{18} to O^{16} ratio which differs from that of another mineral. Then, as the mineralogy progressively changes (as from gabbro to granophyre), the O^{18} to O^{16} ratio will also progressively change.

Another possibility for enrichment of O^{18} in the residuum during fractional crystallization is through progressive assimilation of sediments. Silverman finds that sediments universally are higher in O^{18}/O^{16} than basic igneous rocks (Table 7). Thus, if during crystallization of a basic melt more and more sediment is assimilated, the O^{18}/O^{16} ratio in the residuum will be increased.[7]

Therefore, upon considering the cause of the oxygen isotopic variations without additional data, we must conclude that processes involved in fractional crystallization are as plausible explanations as processes of metasomatism and incomplete mixing.

TABLE 7.—OXYGEN ISOTOPIC COMPOSITIONS IN SILICATE ROCKS

After Silverman (1952, p. 28)

Sample	O^{18}/O^{16} 0/100
Hawaii Sea Water	0.0
Gabbro, No. W-187	6.1
Columbia River basalt	6.4
Disco basalt	6.6
Mt. Etna lava	6.7
Dunite	6.8
Olivine basalt	7.0
Yellowstone obsidian	7.7
Nepheline syenite	8.3
Granophyre, No. W-137-1	9.0
Granophyre, No. W-136-4	9.9
Cheshire orthoquartzite	10.2
Arkose	11.4
Quartz vein	14.0
Grafton orthoquartzite	14.7
Maquoketa shale	15.5
Diatomite	30.3

Alkali Determinations

Alkali determinations were made on five granophyre specimens: (1) to obtain approximate compositions of the microperthite and (2) for comparison purposes. Four of the analyses were determined spectroscopically by Mr. O. Joensu. A fifth is available from the chemical analyses (Table 4). These alkali determinations appear in Table 8.

To determine approximate feldspar compositions, Rosiwal analyses of the specimens were first made. These volume analyses were then recalculated in terms of weight per cent by using the assumed densities given in Table 9. (This table illustrates the leucocratic character of the rock in that four of the five analyses show feldspar and quartz composing over 90 per cent of the rock by weight. The other analysis is over 85 per cent quartz and feldspar). Next, the requisite amount of K_2O was allotted from the alkali determination for the weight per cent of biotite found in the modal analysis. The remaining amount of K_2O and Na_2O was then converted into equivalent amounts of

[7] For further discussion of isotopic variations resulting from fractional crystallization, see Silverman (1952).

orthoclase and albite respectively. These amounts are shown in Table 8 and give approximately the relative amounts of orthoclase and albite found in the microperthite. The total amount of feldspar calculated in this manner agrees well with that found in the mode expressed in weight per cent (*compare* total feldspar content in Tables 8 and 9). This agreement indicates that the calculated composition of the microperthite is essentially correct. However, lime has been neglected in the calculations producing a slight error. The chemical analysis of W-687-1 reveals only 0.2 per cent CaO, and this is believed to be fairly representative of most granophyre specimens. Hence, the error due to neglecting lime will be small.

For purposes of comparison, the above calculations were carried still further. It was observed during thin-section examination that the granophyre essentially belongs to a three-mineral-component system—*i.e.*, quartz, albite, and orthoclase. Thus, with the data from Rosiwal analyses in terms of weight per cent and the relative amounts of orthoclase to albite calculated as described above, the three components—quartz, albite, and orthoclase—were recalculated to 100 per cent (Table 8)

TABLE 8.—ALKALI DETERMINATIONS AND RESULTING MICROPERTHITE COMPOSITIONS PLUS RELATIVE AMOUNTS OF QUARTZ, ALBITE, AND ORTHOCLASE IN GRANOPHYRES

Sample No.*	Alkalies (wt. %)		Calculated feldspar† (wt. %)			Quartz	Albite	Orthoclase‡
	Na_2O	K_2O	Albite	Orthoclase	Total	(wt. %)		
W-177-0	4.2	6.4	35.6	35.3	70.9	28.3	36.0	35.7
W-383	3.3	5.1	28.0	29.2	57.2	41.1	28.8	30.1
W-682	3.4	6.0	28.8	35.1	63.9	33.8	29.8	36.4
W-687-1	4.3	3.2	36.4	18.9	55.3	37.5	21.3	41.2
W1-81	4.0	4.0	33.9	20.1	54.0	38.2	38.8	23.0

* Alkali determinations for W-687-1 were made by H. B. Wiik, Helsinki. The remainder were determined spectroscopically by O. Joensu, Chicago.

† The amount of orthoclase was calculated from the experimentally determined K_2O content only after the correct amount of K_2O had been allotted for biotite (text).

‡ The relative amounts of quartz, albite, and orthoclase have been recalculated to 100 per cent from the amounts found in Table 9. The total amount of feldspar in Table 11 was first converted into orthoclase and albite by using the ratio of Or:Ab indicated in Table 8.

TABLE 9.—ROSIWAL ANALYSES OF GRANOPHYRES EXPRESSED IN TERMS OF WEIGHT PER CENT
Densities taken from Birch (1942, p. 8–15)

Minerals	Volume %	Assumed density	Calculated wt. %	
Granophyre No. W-177-0				
Quartz	26.7	2.654	26.3	93.1
Microperthite	69.0	2.60	66.8	
Mt. + Hem.	2.3	5.15	4.3	
Biotite	2.0	3.40	2.6	
Granophyre No. W-383				
Quartz	38.7	2.654	38.5	93.7
Microperthite	56.8	2.60	55.2	
Mt. + Hem.	1.3	5.15	2.5	
Biotite	1.1	3.40	1.4	
Fluorite	1.7	3.18	2.0	
Calcite	0.3	2.717	0.3	
Miscellaneous	0.1	3.50	0.1	
Granophyre No. W-682				
Quartz	32.0	2.654	31.4	92.9
Microperthite	64.1	2.60	61.5	
Mt. + Hem.	3.3	5.15	6.3	
Biotite	0.4	3.40	0.5	
Miscellaneous	0.2	3.72	0.3	
Granophyre No. W-687-1				
Quartz	34.9	2.654	33.8	90.1
Microperthite	58.2	2.60	55.2	
Albite	1.2	2.63	1.1	
Mt. + Hem.	4.7	5.15	8.8	
Chlorite	1.0	2.8	1.1	
Granophyre No. W1-81				
Quartz	34.5	2.654	32.9	86.2
Microperthite	50.7	2.60	47.4	
Albite	6.2	2.63	5.9	
Mt. + Hem.	5.1	5.15	9.5	
Biotite	2.9	3.40	3.5	
Miscellaneous	0.6	3.83	0.8	

and plotted in Figure 11. This figure is a modification of the equilibrium diagram of the system $NaAlSiO_4$-$KAlSiO_4$-SiO_2 with the region of low temperatures indicated by Bowen (1937, p. 12). Bowen's low-temperature field has been modified in accordance with the more recent work of Schairer (1950, p. 514). In Figure 11 the low-temperature field is bounded by two 1075° C. contours and the cotectic line joining the quartz-feldspar fields.

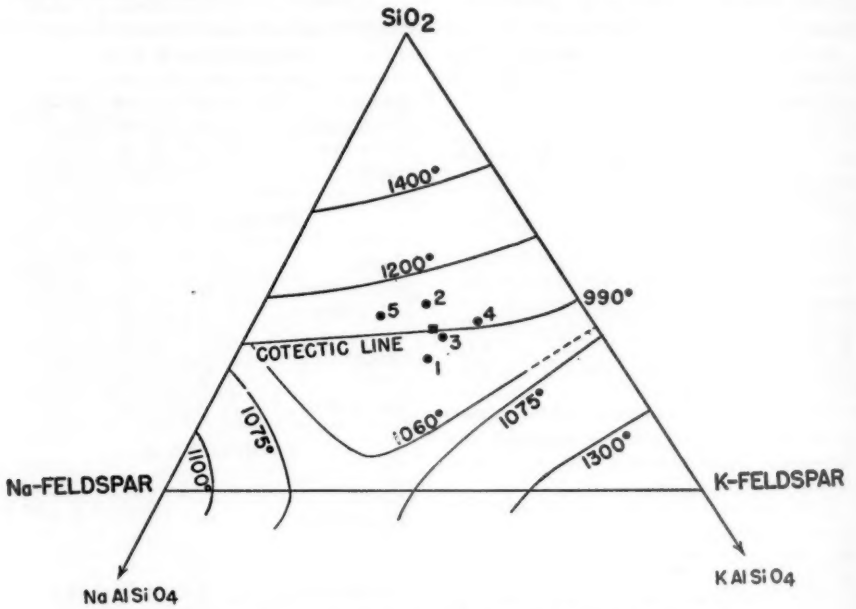

FIGURE 11.—COMPOSITION-TEMPERATURE RELATIONS OF THE GRANOPHYRE

Points labelled 1, 2, 3, 4, and 5 represent specimens W-177-0, W-323, W-682, W-687-1, and W1-81, respectively. The square on the cotectic line is the average of these five specimens. Diagram modified after Schairer (1950, p. 514).

GEOLOGICAL INTERPRETATION

Origin of Crystalline Rocks in the Complex

Gabbro.—Throughout this paper, the gabbro has been assumed to be of magmatic origin. This interpretation has recently been questioned for gabbros in similar complexes. For example, Reynolds (1947b, p. 222) states that "granitization and the complementary basification are major petrogenetic processes adequate to account for associations which suggest the term 'contrasted differentiation'."[8] Two years later, van Biljohn (1949) suggested that the layered Bushveld norite as well as the associated granophyres and granites resulted from transformation (metasomatism) of sediments.

That the Wisconsin gabbro cannot be a product of metasomatism is shown by the correspondence of its O^{18}/O^{16} ratio with true basic igneous rocks (Table 7). Further criteria of a magmatic origin are: (1) the feldspars of the gabbro commonly exhibit oscillatory zoning—a characteristic of crystallization in a magma; (2) in contact with the Davis Hill clastics and with the Huronian formations in the southern part of the area, the gabbro shows chilled contacts; and (3) the gabbro is associated with basic flows and diabase dikes which have identical mineral compositions and tex-

[8] Holmes (1936) referred to gabbro-granophyre complexes as showing "contrasted differentiation".

tural relations. The gabbro meets exactly the requirements given by W. Q. Kennedy (1933) for the tholeiitic rock type developed from a basaltic magma.

Intermediate zone.—The narrow zone between the gabbro and granophyre was investigated in detail both microscopically and chemically. Chemical investigation failed to distinguish between a primary or secondary origin for the intermediate rock. However, field and microscopic examination has revealed 14 significant facts for the interpretation of origin:

(1) In places the intermediate rock shows a pseudo-diabasic texture which resembles the true diabasic texture of the gabbro.
(2) The plagioclase crystals show a peculiar irregular embayed type of zoning.
(3) Anastomosing stringers of acid feldspar cut more basic cores.
(4) A blue-green amphibole occurs in felty aggregates and in ragged patchy crystals which show variable optical properties.
(5) "Ghost" remnants of large plagioclase crystals are indicated by sericitic patches extending across aggregates of microperthite crystals.
(6) Feldspar shows considerable alteration to epidote, sericite, and chlorite.
(7) Ghostlike remnants of mafics, now cryptocrystalline yellowish masses, appear in feldspar crystals.
(8) Quartz replaces feldspar in places.
(9) Oriented ragged pyroxene relicts occur in feldspar.
(10) Oriented ragged amphibole relicts occur in feldspar.
(11) Hornblende is pseudomorphous after pyroxene.
(12) The intermediate zone occurs between flows and red rock as well as between gabbro and red rock.
(13) Apatite is particularly abundant in the intermediate rock.
(14) Tourmaline occurs in this zone.

In the field, the pseudo-diabasic texture of the intermediate rock becomes more prominent near the gabbro. Under the microscope, the texture is due to the association of amphibole and plagioclase rather than to pyroxene and plagioclase. With changing mineralogy the pseudo-diabasic texture grades into the true diabasic texture in the gabbro. Hence, this texture is best regarded as pseudomorphous after the diabasic texture of gabbro.

The peculiar, irregular, embayed type of zoning is a common phenomenon in the intermediate zone (Pl. 3, fig. 1) and does not resemble the evenly distributed zones so characteristic of most igneous rocks. Instead, it has the appearance of corrosive action. In many places, this zoning is related to the next described feature.

The occurrence of anastomosing stringers of albite cutting through more basic cores illustrates the action of sodic material when in non-equilibrium with the rock (Pl. 3, fig. 2). Gilluly (1933, p. 68) found the same phenomenon of albite stringers in andesine in the metasomatic albite granite near Sparta, Oregon.

The blotchy appearance of single amphibole crystals with their concurrent variability in optical properties is hardly suggestive of primary origin. The blue-green fringe on many amphibole crystals, and irregular blue-green patches in others, suggest introduction of soda. Also, the occurrence in felty aggregates, and irregular, corroded crystals suggests alteration processes.

Ghostlike remnants of plagioclase indicated by rectangular sericitic patches extending across several anhedral microperthite grains suggests that the rock was originally of different character. In addition to this feature, much of the plagioclase in the intermediate zone shows considerable epidote and white mica alteration. The lime in the epidote suggests that the plagioclase was formerly more calcic.

Patches and elongated masses of yellow, minutely crystalline material are embedded in feldspar crystals in places and appear as ghostlike remnants or relicts of some former ferromagnesian mineral.

Introduction of silica is indicated by the quartz replacement of feldspar (Pl. 3, fig. 4). The evidence of quartz replacement here in the intermediate zone may indicate that the granophyric interstices in this rock also originated through replacement.

Oriented pyroxene and amphibole crystals imbedded in plagioclase indicate corrosion and replacement by feldspar (Fig. 8; Pl. 3, fig. 3).

Alteration of pyroxene has produced am-

phibole pseudomorphs. Pyroxene parting and patchwork resembling the "graphic" intergrowth of hypersthene and augite found in the gabbro are preserved.

The intermediate zone between flows and red rock suggests secondary origin. At places the flows and gabbro have a single red rock mass in common. As the flows are oldest, they surely did not differentiate simultaneously with the younger gabbro to produce the intermediate zone and subsequently the single mass of red rock. Hence, if the narrow intermediate zone adjacent to flows is secondary, then the larger one between gabbro and red rock may also be secondary.

The abundance of apatite in the intermediate zone indicates alteration (Pl. 3, fig. 5). One specimen of intermediate rock showed 3.6 per cent apatite. Nockolds (1932, p. 442), in describing the alteration of basic xenoliths in granite, noted the abundance of apatite needles. He found that

> "the apatite needles are scattered all through the feldspars in the xenoliths and are not found only in the margins of the crystals. This shows that the whole feldspar has been altered and not merely the margin. . . ."

One year later, Nockolds (1933) cited 11 examples of the concentration of apatite in basic rocks intruded or included by acidic magmas and believed that this concentration resulted from volatile action.

Tourmaline, found in one sequence of rocks in the intermediate zone, is apparently out of place if this zone is of primary origin. That is, tourmaline ordinarily occurs as a primary mineral in granitic or pegmatitic rocks. Hence, its presence in the relatively basic intermediate zone suggests later introduction of material.

The 14 features just described agree with a hypothesis of secondary origin. As pointed out by Schwartz and Sandberg (1940, p. 1149), gabbro and red rock

> ". . . grade into each other, but it is significant that the gradation is very largely a result of association of the two main types of material rather than gradation through diorite, syenite, and other types."

This failure to find ordinary differentiates rather than the peculiar intermediate rock argues against differentiation. On the other hand, the complete gradation in texture, and the sequence in alteration of both ferromagnesian minerals and feldspars from red rock into gabbro, plus the 14 features previously discussed constitute overwhelming evidence that the intermediate zone originated through alteration of gabbro. Just as the intermediate zone between gabbro and red rock is due to alteration of gabbro, the intermediate zone between flows and red rock is due to alteration of flows.

What was the cause of this alteration? The answer to this problem lies in the origin of the granophyre.

Granophyre.—The major problem of the area concerns the origin of the granophyre. Did it originate by metasomatism, or did it result from solidification of a melt?

Mineralogical features in the granophyre, such as the presence of chlorite and kaolinite, the corroded outlines of biotite, the chessboard appearance of albite, and the high content of albite in plagioclase suggest secondary processes. In addition, there are field evidences of secondary processes—*e.g.*, the presence of altered albitized basic flows and the existence of an alteration zone between gabbro and granophyre. But are these replacement features the result of metasomatic action forming granophyre or the result of magmatic action of a granitic melt? The criteria for a replacement origin of the granophyre are inconclusive.

On the other hand, let us consider the interpretation of these 11 facts:

(1) Rotated blocks of gabbro are present in the granophyre.
(2) From granophyre to gabbro, grain size decreases in the area of the rotated gabbro blocks.
(3) In one red rock dike cutting the gabbro, alkali feldspars are oriented parallel to the contact and perpendicular to the structure in the gabbro.
(4) Another red rock dike cutting diagonally across the fluxion structure of gabbro is massive. There is no preservation of gabbro structure.
(5) Granophyric interstices are present in acid flows.
(6) The granophyre possesses miarolitic cavities.

(7) Dilationlike features are present on both a microscopic and macroscopic scale.
(8) One granophyric spherulitic red rock dike shows an aphanitic groundmass.
(9) A plot of analyses in the system quartz-albite-orthoclase shows a clustering of points along the cotectic line.
(10) Analyses of alkali ratios show that microperthite has a ratio of Or:Ab close to 50:50.
(11) Acid igneous activity occurred both before and after granophyre formation.

Each of these facts will be discussed in the order given.

Discussion of 1 and 2: The presence of rotated blocks of gabbro in granophyre (fig. 1) indicates a mechanical rotation. However, whether these fragments represent skialiths (Goodspeed, 1948b, p. 515) derived from a fault breccia or true xenoliths is debatable. If these fragments are remnants of a fault breccia, then the present granophyric matrix resulted from replacement of the breccia matrix. At Cornucopia, Oregon, Goodspeed and Fuller (1944, p. 274) found that finely crushed material in a tectonic breccia was more susceptible to replacement than the larger fragments, and thus here, as at Cornucopia, the larger fragments may represent material less favorable for replacement than the matrix.

In a series of articles, Goodspeed (1944; 1948a; 1948b) discusses the criteria for relict fragments in a replacement breccia. He states (1948b, p. 520):

> "Enclosed by granitized material, the fragments of a replacement breccia are usually very irregular in size and shape; some consist of relatively large pieces joined by thin links. The contacts of the fragments and the matrix usually exhibit crystalloblastic textures and may be gradational. The minerals of the matrix are not only crystalloblastic but the overall texture is typically irregular and uneven. Minerals similar to the matrix appear as porphyroblasts in the fragments, and crystalloblastic invasion along joint cracks is common. . . . As granitization proceeds, the relict fragments are changed in their mineral composition and texture but retain an angular shadowy outline that fully warrants the term skialith."

Again (p. 522) Goodspeed says: "It is noteworthy that skialiths exhibiting various kinds and degrees of mineralogic and textural changes may be found within a few inches of each other." Another pertinent statement (p. 524) reads: "Although relatively small skialiths are most abundant along border zones of granitized masses, larger relict fragments are to be found well within a granitic mass." Another criterion used by Goodspeed (1948a, p. 66, 69, 72, 74) is the presence of slickensides and other cataclastic structures.

A comparison of the present breccia zone with those described by Goodspeed reveals little correspondence in features. Contacts of the gabbro blocks and the matrix are relatively sharply defined. The texture of the matrix is regular and even. Lacking are the development of porphyroblasts in the fragments, the crystalloblastic invasion along joint cracks or along the planes defined by the fluxion structure, the ghostlike remnants or fragments with shadowy outline, the varying kinds of mineralogic and textural change from one fragment to another, and the presence of slickensides so common in those occurrences cited by Goodspeed. Everywhere the fragments are gabbroic and are identical to the main gabbro mass a few hundred feet to the north. Noteworthy is the absence of fragments in any other area except this one small zone (not over a few hundred square feet) in sec. 29, T. 45 N., R. 4 W. This patch of breccia hardly has the shape or extent of a fault zone. Definite walls are absent. In this respect, Goodspeed (1948a, p. 72) states: "The lack of definite walls to this breccia zone [one] at Cornucopia as well as the absence of slickensides and gouge rules out any interpretation of past . . . faulting"; and, therefore, he concludes the breccia is a flow breccia. Thus, the spatial relations of the breccia zone and the lack of replacement criteria argue against an interpretation of fault-breccia replacement. Instead, the best interpretation appears to be one in which the fragments represent xenoliths which moved downward only a short distance before being trapped in the chilled border of the main red rock body. The granophyre shows a decrease in grain size toward the breccia zone and the gabbro, as if the granitic melt chilled rapidly adjacent to the cooler, solidified gabbro.

Discussion of 3: It would be a remarkable coincidence indeed if replacement were to produce the following five coexisting features found in one red rock dike cutting gabbro:

(1) the red rock dike possesses a massive center; (2) grain-size decreases from the center to the margin; (3) parallel orientation of small alkali feldspar laths is found along the margin of the dike; (4) the orientation of these feldspars parallels the contact; and (5) the orientation is perpendicular to the lineation in the country rock—the country rock in this case is of the intermediate rock type which grades imperceptibly into gabbro within a few feet of the contact. The interpretation of these five features agrees better with the hypothesis of magmatic injection in which chilling produced the decrease in grain size, and frictional drag along the walls produced the feldspar orientation.

Discussion of 4: Another red rock dike, which cuts diagonally across the fluxion structure of gabbro, is massive. This fact can be used as a negative criterion. Inasmuch as replacement products commonly preserve the structure of pre-existing rocks, the massive structureless crosscutting dike suggests a nonreplacement origin.

Discussion of 5: In a few places—*e.g.*, the south of sec. 19, T. 45 N., R. 4 W., and in some felsite fragments in the clastic rock of Davis Hill—acidic igneous flows actually possess granophyric interstices. One felsite outcrop in sec. 19, although discontinuous, shows an apparent increase in grain size toward the core where it becomes granophyric. Surely, if granophyric texture occurs in flows in the area, there must be some subsurface counterpart.

Discussion of 6: The red rock contains miarolitic cavities lined by quartz, fluorite, and, in places, biotite. Regarding such structures, Goodspeed (1948a, p. 62) states:

> "Although more visible effects of volatiles during the magmatic stage of formation are much more numerous in surface lavas or magmas crystallizing near the surface, yet the presence of miarolitic or smaller miniature miarolitic-like cavities can hardly be explained in any other way than due to the escape of volatiles."

Thus, the formation of such structures is probably related to magmatic action.

Discussion of 7: Figure 3 shows dilationlike features in a composite dike. In thin sections of the intermediate zone, microdilation features include a quartz vein offsetting an epidote stringer, and another quartz vein intersecting and offsetting a plagioclase crystal. In all three examples the position of the offsets suggests magmatic action (Goodspeed, 1940, p. 194). However, these dilation dikes and veins could be secretion dikes formed in the manner suggested by Ramberg (1949, p. 50).

Discussion of 8: The pertinent features of a porphyritic, granophyric, spherulitic, red rock dike cutting felsite flows in sec. 30, T. 45 N., R. 4 W., were described in the section of Petrographic Data (Pl. 2, fig. 5). The texture of this dike is typical of igneous rocks; resorbed dipyramidal quartz and microperthite phenocrysts rimmed in places by a radiating spherulitic growth of feldspar or feldsapr and quartz are imbedded in an aphanitic groundmass with grain size averaging less than .014 mm. Such phenomena are duplicated in volcanic areas. For example, the writer examined a thin section[9] of an acidic dike associated with the Tertiary volcanic activity in the Black Range, New Mexico, and found identical features. Indeed, the similarity of the two dike rocks is striking.

This red rock dike bridges the gap between the granophyre and the felsites of the area—*i.e.*, the phenocrysts of quartz and microperthite and the intergrowth of quartz and feldspar in the spherulitic parts resemble the granophyre, whereas the aphanitic groundmass is identical to that in the acidic felsite flows. Hence, this dike records the different stages of cooling, leading first to granophyre production and then to extrusive felsite formation. Quartz and microperthite represent the earlier stage of slow cooling at depth; the spherulites represent moderate undercooling perhaps accompanying the intrusion into its present locale; and the aphanitic groundmass, rapid cooling and chilling by a colder country rock.

Discussion of 9: A plot of analyses of five red rocks in the system quartz-albite-orthoclase shows a clustering of points along the cotectic line (Fig. 11). The red rock is essentially a three-mineral-component system; therefore this clustering of points along the cotectic is to be expected if the rock formed from a melt. Would metasomatism also give this result?

[9] This thin section was kindly supplied by Dr. F. J. Kuellmer of the New Mexico Bureau of Mines.

Bowen (1950, p. 655) believes that "a replacement process of any kind should be completely lacking in any tendency to produce an association of minerals in definite or approximately definite proportions." In continuation of Bowen's argument, why should metasomatism give minerals in such a proportion that they tend to cluster around the cotectic line? Is this not a criterion of magmatic action? Also the average composition of the five granophyres falls extremely close to the cotectic line.

The deviations in Figure 11 may be explained on the basis that the system is not a pure three-component-system and that the cotectic line was derived through laboratory study on a dry melt at atmospheric pressure. Hence, the addition of pressure or water or other impurities may shift the cotectic line.

Discussion of 10: In the five analyzed red rock specimens, the amount of albite in the microperthite ranges from 45 to 66 per cent of the total feldspar (Table 8). If plotted in the equilibrium diagrams for the alkali feldspars given by Bowen and Tuttle (1950, p. 496), these compositions fall along the rather flat portion of the solidus curve close to the minimum melting point. A variation of less than 10° C. would account for all the differences in composition between the analyzed specimens—*i.e.*, if they had formed from a melt in the three-component-system $NaAlSi_3O_8$-$KAlSi_3O_8$-H_2O. The fact that the composition of these microperthites is near the center of the low melting valley in this equilibrium diagram may indicate that these so-called residua of a magma actually resulted from magmatic crystallization. Tuttle (1952) finds that a frequency-distribution curve of the composition of perthites in extrusive rocks shows a maximum between Ab_{60} and Ab_{80}. This suggests that magmatic action is responsible for most perthites of this composition. Hence, since the perthites in the present rock show an average between 50 and 60 per cent albite, they too may have resulted from magmatic action. A true test of Tuttle's hypothesis of magmatic origin for perthites of this composition would be to establish a distribution curve for known metasomatic perthites and then see if any difference really exists in the maximum between the magmatic rocks and the metamorphic-metasomatic rocks.

Discussion of 11: Acidic igneous activity during the Keweenawan occurred both before and after the formation of granophyre. The earlier igneous activity is shown by the presence of felsite and rhyolite flows which are cut by red rock dikes; whereas, later igneous activity is illustrated by an aphanitic felsite porphyry which shows flow laminae and which cuts both granophyre and gabbro. Since acidic igneous activity was present in this region both before and after granophyre formation, the granophyre was probably formed by the same igneous activity. Furthermore, if acidic flows are present in the area, where are their subsurface counterparts? Here, we have a simple and logical answer—the granophyre. Lavas must signify magmas at depth.

Origin of Acidic Magma

Several possibilities for the mode of origin of the granophyric melt have been suggested: (1) assimilation or syntexis, (2) liquid immiscibility, (3) intrusion from a primary granitic magma, (4) differentiation *in situ* by fractional crystallization, (5) separate intrusion due to filter-press action, and (6) refusion.

Assimilation.—Whether assimilation played a part in the development of red rock magma is difficult to prove or disprove. Bowen (1928, p. 184) argued that magmas do not possess any degree of superheat and consequently cannot assimilate any large quantities of country rock. On the other hand, Daly (1917; 1933) and Fenner (1931; 1937) have been strong advocates of the assimilation hypothesis. In the present area—at least in the southern gabbro belt—basic magma could not have assimilated any large quantities of acidic material, because over a large part of this area the gabbro is in contact with basic flows. In addition, there is no trace of xenoliths of acidic rock in the gabbro. Tyrrell (1928, p. 146) states that ". . . it is clear that assimilation and hybridization effects should only be inferred when there is evidence of magmatic mingling or the dissolution of xenoliths." The effect of assimilation in producing the variations in the oxygen isotopic ratios was previously discussed. The writer agrees with Schwartz and Sandberg (1940, p. 1166) who, in their study of Duluth sills, con-

cluded that assimilation could not be excluded as a process, but thought it improbable that it was the dominant process.

Liquid immiscibility.—The improbability of liquid immiscibility in melts of ordinary rock compositions was demonstrated by Greig (1927). However, in two recent articles (Fenner, 1948; Barth and Rosenquist, 1949) the hypothesis of liquid immiscibility has been revived. Both state that liquid immiscibility may exist in magmas above the temperatures of crystallization. However, with cooling, and at ordinary temperatures of crystallization, the two immiscible fractions recombine into one conjugate liquid. If convection currents have played a role during the crystallization of gabbro—*e.g.*, in such a manner as suggested by Grout (1918a; 1920) and Wager and Deer (1939)—then this remixing should be fairly complete. However, if remixing has not been complete, then along the zone of intermingling —*viz.*, the intermediate zone—there should be a systematic variation of composition, but chemical analyses along this zone show geochemical culminations. In addition, the spatial relations of gabbro and granophyre argue against an origin through segregation of immiscible fractions. The lighter immiscible fraction should occur above the heavier. Instead, the gabbro-granophyre complex shows an intertonguing of the two rock types. Therefore, such an occurrence as red rock and gabbro with but a narrow transition zone is not likely to be a result of two immiscible liquids.

Primary granitic magma.—Several writers have suggested that primary granitic magma exists at depths. This hypothetical source for a granophyre magma is eliminated by the nearly constant association of granophyre and gabbro throughout the world.

Crystallization-differentiation in situ.—Whether or not differentiation by fractional crystallization will produce a granitic residua is a much debated question. In this area the hypothesis of differentiation in place through fractional crystallization meets several objections.

The red rock is decidedly younger than the gabbro since such features as composite dikes, intrusion breccias, and common dikes occur. It is difficult to believe that these features represent auto-intrusion phenomena. In one composite dike, red rock penetrates a basalt dike, the latter showing chilled contacts with the gabbro. Hence, the red rock was introduced after the gabbro was sufficiently cool to give the chilled contact effect. Time for complete solidification of gabbro is also illustrated by intrusion breccia and many red rock dikes cutting gabbro.

The amount of red rock in proportion to the amount of gabbro is another argument against differentiation *in situ*. Grout (1926, p. 549) calculated that a maximum of one-tenth of an average basalt magma may become granite. Plate 1 shows a considerably larger ratio of granitic rocks to gabbroic rocks than can be accounted for through simple differentiation *in situ*. Also intermediate rocks are very scarce.

The shape of the red rock bodies contradicts the hypothesis of differentiation in place. Arms of red rock projecting into gabbro are certainly not local accumulations of residua produced through the action of crystal settling. In addition, the red rock mass in secs. 9, 10, 15, and 16, T. 44 N., R. 5 W., shows no floor or roof of gabbro and, thus, appears unrelated to gabbro in these parts.

The presence of an alteration zone between gabbro and granophyre does not support differentiation in place. It is more compatible with an interpretation of a later intrusion of a granitic magma which produced contact alteration of its wall rocks. In places the break in textural relations and mineral proportions is sharp between red rock and intermediate rock, but intermediate rock grades into gabbro. This suggests contact action of granophyre on gabbro.

Separate intrusion through filter-press action.—A residua formed by fractional crystallization at depth may have been squeezed out by filter-press action and injected into its present locale where it solidified as granophyre. This hypothesis answers all the objections to the hypothesis of differentiation *in situ*, but it may be doubted whether the hypothesis of separate intrusion through filter-press action will give the nearly constant association of gabbro and granophyre found throughout the world.

Refusion or palingenesis.—Read (1943, p. 79) states that remelting may occur by two methods.

> "In the first place, depression of the sial into the substratum by movements during initiation and growth of the fold-belts of the crust may lead to its pure melting," and secondly, "instead of depressing the sial down into the substratum to be melted, we could consider the substratum to rise into the upper layers, bringing sufficient heat with it to melt important masses of the sial with the formation of granitic magmas."

With regard to the first method, both Anderson (1938) and Buddington (1943) believe that magmas can form through some manner of raising the temperature-depth curve relative to the fusion curve for a particular kind of rock. Hence, in this area the granophyre magma may have originated during the basining of the Lake Superior geosyncline. That is, rocks of appropriate composition in the lower reaches of the geoxyncline were brought to a point of fusion or partial melting, and the resulting liquid squeezed out into the limbs of the syncline where it followed the same planes of weakness which earlier influenced the gabbro injections.

With regard to the second method suggested by Read, the heat from the intrusion of gabbro could scarcely have produced such a volume of granophyre in the present area, partly because gabbro was intruded into basic flows over the greater part of the area (and, thus, refusion would produce only additional basic magma), and partly because of the apparent lack of superheat of basic intrusions in general (Bowen, 1928).

Summary of modes of origin of red rock magma.—Considering only the present area, it seems that a separate intrusion originating from: (1) filter-press action of a differentiate at depth, (2) refusion at depth and squeezing upward of the derived liquid, or (3) a primary granitic magma offers the best explanation of the observed phenomena. These phenomena include the alteration zone between gabbro and granophyre and the younger age of granophyre as shown by diking of gabbro, the formation of composite dikes, and the unoriented xenoliths of gabbro in red rock. However, if other gabbro-granophyre areas are considered, it seems unlikely that the constant association of granophyre and gabbro may be explained through intrusion from a primary granitic magma. Refusion and the filter-press mechanisms are also questionable in the light of this genetic relationship. However, the structure that controlled gabbro intrusion may also have exerted its influence on granitic magmas generated by refusion or crystal differentiation at depth.

Mechanics of Formation of Intrusive Rocks

Gabbro.—The intrusion of gabbro shows definite relations to the structure of the region. Torsional warping as a result of differential foundering has been described. The stress produced by this torsional warping resulted in slippage between flow units and in thrust faults. Along these planes of weakness gabbro was intruded.

Granophyre.—The torsional warping also produced two oblique structures, an anticline and a syncline cutting diagonally across the northward-dipping limb of the Lake Superior geosyncline. If movement along planes of shear followed gabbro intrusion, "openings" or areas of low pressure would occur between these two oblique structures, and thus would be favorable structural areas for intrusion of granophyre magma (Fig. 4). This hypothetical picture, therefore, offers an explanation for the belted character of the complex.

Intermediate zone.—This zone is believed to have acquired its peculiar alteration character from contact action of the intruded granophyre magma on the solidified gabbro. Perhaps both mechanical and chemical processes were responsible for the odd mineralogical and textural features of this transition zone. Mechanical processes are indicated by fracture fillings in one cataclastic area of the transition zone. In addition, some of the cores of albite (xenocrysts?) in granophyre close to the granophyre-intermediate rock contact may indicate some disintegration of gabbro. Hence, some mixing due to mechanical processes is indicated.

For the most part, however, the absence of mechanical features indicates that introduction of material has been through chemical processes, *i.e.*, diffusion. The driving forces of these

diffusion processes are differences in chemical activity (Ramberg, 1945, p. 99) or chemical potential (Bugge, 1945, p. 57), which, in turn depend upon such variables as temperature, pressure, concentration, and the gravitational field. Here, gabbro and granophyre differ greatly in composition, and we can conclude that these differences were in large part responsible for the chemical activity gradients. Ramberg (1945, p. 106) believes that the most important processes of contact metasomatism are caused by the difference in composition of the surrounding rocks and of the gases and solutions from the magma masses. In addition, a temperature gradient between the hot granophyre magma and the solidified country rock probably existed. These two factors, temperature and composition, are regarded, therefore, as the greatest factors in the chemical activity gradients. Pressure differences were probably negligible, and the effect of the gravitational field was not great because granophyre was intruded after the rock had been at least partially tilted to the north; thus, the variation from granophyre to gabbro was produced essentially horizontally. It is concluded that the oxygen isotope variations in the transition zone resulted from metasomatic processes and not from fractional crystallization. Similarly the geochemical culminations were produced through metasomatic processes.

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EXTENSION OF TAZEWELL GLACIAL SUBSTAGE OF WESTERN ILLINOIS INTO EASTERN IOWA

By Paul R. Shaffer

Abstract

The Green River lobe of Tazewell (Shelbyville) ice extended west of the Mississippi River to the eastern border of the Goose Lake Channel in Clinton County, Iowa. It extended northwestward to within 1½ miles of the village of Goose Lake, across an area previously mapped as Iowan and Kansan drift. A small area of Illinoian drift is present between the Shelbyville drift and the Goose Lake Channel.

The Iowan drift is separated from the western border of the Goose Lake Channel and the Shelbyville drift by a small strip of Kansan drift on the north and a small area of Illinoian drift on the south.

When the Shelbyville ice reached the "big bend" of the Illinois River in Bureau County, Illinois, it blocked the eastward-flowing drainage of the old Mississippi River and formed the first stage of glacial Lake Milan which extended from the "big bend" to the rock-defended outlet near Andalusia, Illinois. Later when the Shelbyville ice had advanced westward to the west side of the present Rock River valley, glacial Lake Milan was restricted to the lower Rock River valley, and glacial Lake Cordova formed north and northwest of the ice dam. The outlet for glacial Lake Cordova was over a divide composed of Illinoian drift between Cordova, Illinois, and Le Claire, Iowa. Later when the Shelbyville ice crossed the present Mississippi River valley, glacial Lake Savanna was formed to the north. Its waters rose until they found an outlet through the Goose Lake Channel to the Wapsipinicon River, down the Wapsipinicon and through the new valley past Cordova and Le Claire. When the Shelbyville ice withdrew, the present course of the Mississippi River between Fulton, Illinois, and Keokuk, Iowa, had been established.

Two sets of loess-covered terrace remnants, deposited under different controlling conditions, are present between Geneseo and Lomax, Illinois. One set between Geneseo and Andalusia, Illinois, rises to a nearly uniform height and resulted from quiet-water deposition in glacial Lake Milan. Below Muscatine, Iowa, lie remnants of deposits made up largely of fine sands and silts, whose deposition was controlled by a constriction in the valley of the Mississippi River near Keokuk. When the great volume of water from the upper Mississippi drainage was added to that already flowing past Keokuk, constriction caused higher water levels upstream with areas of relatively quiet water in marginal portions of the valley and in the ponded tributaries.

The possible contemporaneity of the Shelbyville terrace on the lower Iowa with the high terrace of Lake Calvin upstream calls for a re-examination of the terraces along the Iowa River and raises the question of the possibility that the Shelbyville and the Iowan were contemporaneous.

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Introduction

For more than 60 years the invasion of ice from Iowa into northwestern Illinois during the Iowan substage of the Wisconsin stage has been a moot question. Leverett (1899, p. 151) felt that evidence favored the invasion but left the question open for subsequent study. Leighton (1923, p. 267) concluded that "A post-Illinoian ice sheet invaded the Green River basin.... The evidence is insufficient to determine whether this ice lobe, now called the Green River lobe, is Iowan in age or a correlative of the Shelbyville drift of Early Wisconsin age." Subsequent work by Templeton (1939, unpub. thesis) and Ekblaw (1942, p. 154) led them to believe that the Green River lobe was related to the Shelbyville moraine of the Tazewell substage of the Wisconsin stage.

Figure 1.—Index Map
Area studied shaded

In 1948 Leighton examined some new openings in the Garden Plain esker (Leverett, 1899, p. 79) near the bluff of the Mississippi valley. The fresh character of the materials of the esker and the conformable relations of the overlying loess to the gravels, plus evidence of the damming of the Mississippi River found by him at Savanna, Illinois, led to further search in Illinois and Iowa for other evidence of a westward extension of the Green River lobe beyond the tentative boundary proposed by him in 1923. The first report on these studies (Leighton and Shaffer, 1949) was made before the maximum westward and northwestward extent of the Green River lobe was known, and before the Tazewell drainage history of the Mississippi Valley between Clinton and Keokuk, Iowa, had been completely determined.

The area covered by the writer is shown in Figure 1.

Acknowledgments

These studies were made under the auspices of the Illinois State Geological Survey and the Iowa Geological Survey. Dr. M. M. Leighton, Chief, Illinois State Geological Survey, suggested the possibility of a western extension of the Green River lobe and aided greatly in the field work and with the preparation of this report. Dr. H. G. Hershey, Director, Iowa Geological Survey, and Professor A. C. Trowbridge, State University of Iowa, served as consultants in the studies. Their aid, advice, and counsel have been extremely valuable. Dr. H. B. Willman and Dr. George E. Ekblaw of the Illinois State Geological Survey contributed valuable suggestions and data. Professor George W. White, University of Illinois, made helpful suggestions during the preparation of the report. Mr. Herbert G. Stewart, Jr., and Mr. Vincent C. Shepps served as research assistants for one summer. Their aid is gratefully acknowledged.

Evidence for Extension of Shelbyville Ice into Eastern Iowa

Green River Lowland

A lowland which had persisted from preglacial time was present between Fulton, Illinois,

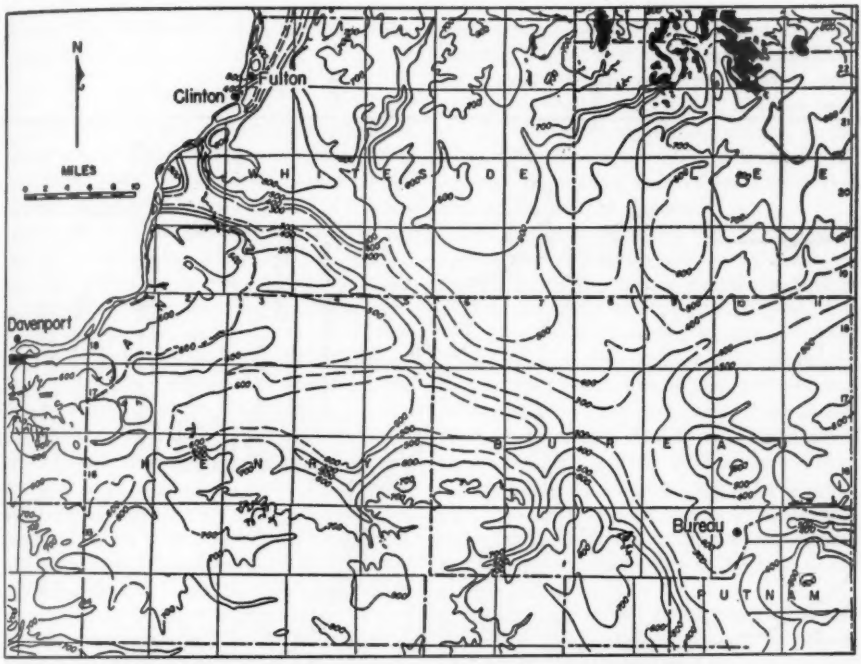


FIGURE 2.—BEDROCK TOPOGRAPHY OF PART OF NORTHWESTERN ILLINOIS

Showing preglacial bedrock valley between Fulton and "big bend" of Illinois River at Bureau, Bureau County. Reproduced from Horberg's "Bedrock Topography of Illinois": Ill. State Geol. Survey Bull. 73.

and the Illinois River during the advance of the Tazewell lobe. The lowland was mentioned by Udden (1899, Pl. VII, p. 322) as a possible course for the preglacial Mississippi River. It extended southeastward from Clinton through the Meredosia Channel and Green River Basin to the "big bend" of the Illinois River near Bureau (Fig. 2).

Horberg's (1946, p. 191) extended work on the bedrock surface and the drainage systems of Illinois has shown that in Sangamon time the Mississippi had a course southeastward from Fulton to what is now the Illinois River in Bureau County, Illinois. Later Horberg (1950, Pl. I, sheet 1) presented a contour map of the bed rock surface of Illinois showing this major bedrock valley (Princeton Valley) from Fulton to the "big bend" of the Illinois River (Fig. 2).

Trowbridge *et al.* (1941, p. 296) and Leverett (1941, p. 1920) had previously recognized that the lake which formed in front of the western margin of the Illinoian ice in Iowa, called Lake Calvin (Fig. 3), was of short duration, and that,

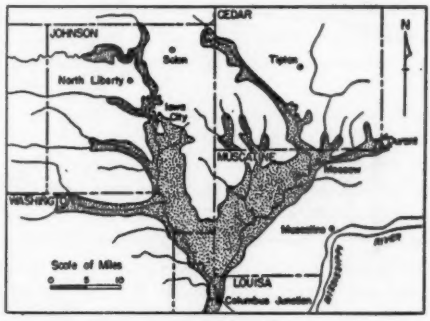
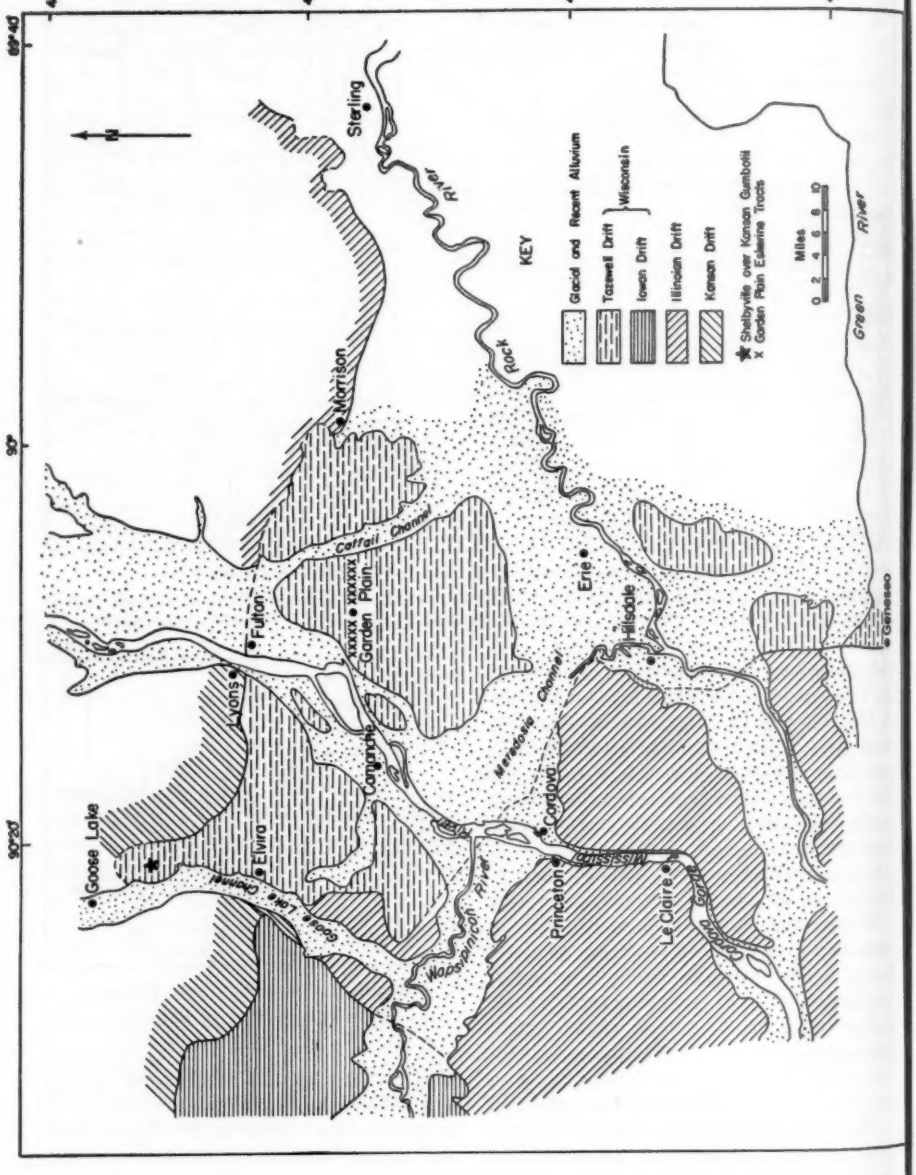


FIGURE 3.—GLACIAL LAKE CALVIN

After Schoewe (1924, pl. VI, p. 56)

following the retreat of the Illinoian ice, the upper Mississippi River drainage returned to its southeastward course to the Illinois River and continued there during the Sangamon interglacial stage.

The presence of this broad valley in post-Sangamon time facilitated the western exten-


42°
41°50'
41°40'
41°30'
90°40'
90°
89°30'
N
Sterling
Rock River
Green River
Morrison
Fulton
Lyons
Elvira
Goose Lake
Goose Lake Channel
Cordova
Princeton
Le Claire
Wapsipinicon River
Mississippi River
Meredosia Channel
Cattagi Channel
Garden Plain
Erie
Hillsdale
Geneseo
KEY
Glacial and Recent Alluvium
Tazewell Drift
Wisconsin
Iowan Drift
Illinoian Drift
Kansan Drift
Shelbyville over Kansan Gumbotil
Gorden Plain Edamite Trocts
Miles
0 2 4 6 8 10

sion of the Shelbyville lobe, also called the Green River lobe.

Garden Plain Esker

Leverett (1899, p. 79) described a gravel-pit exposure in an esker in the western part of sec. 22, Garden Plain Township, Whiteside County, Illinois, about 2½ miles east of its western terminus, and traced the esker westward to the bluff of the Mississippi River (Fig. 4). Leighton observed additional and more recent openings in the esker farther west in sec. 20. Other openings in the esker system are known as far east as Cattail Creek in sec. 19 of Union Grove Township, Whiteside County. The cross-bedding dips westward, indicating deposition by a westward-flowing, subglacial stream. The gravels are extremely fresh where overlain by calcareous and fossiliferous loess. Near the western end, where the gravels are overlain with only a few feet of loess, there is practically no staining, and the limestone pebbles and cobbles are fresh and firm. Near the eastern end of the eskerine tract the gravel, which underlies about 6 feet of sand, is rusty brown, the coloring, however, being the result of ground-water oxidation. Some of the limestone pebbles are partially dissolved, while adjacent ones are fresh. Leverett (1899, p. 79) felt "that the esker must be referred either to the Illinoian or to an extension of the Iowan ice westward from the Rock River Basin." To refer the esker to the Illinoian requires truncation of the Sangamon profile of weathering just prior to loess deposition. The Garden Plain esker, however, lies within an eskerine tract. The esker is regarded as Shelbyville in age, formed by a stream flowing beneath the ice of the Green River lobe, because (1) there is no interval of weathering between the fresh gravels of the esker and the overlying calcareous and sometimes fossiliferous loess; (2) the esker occurs in a region of constructional topography; and (3) till younger than Illinoian occurs in the surrounding area and is similar to Shelbyville till of the Green River lobe farther east.

Older Drifts

Kansan Drift

North, west, and northwest of the Shelbyville drift in Iowa, the Kansan drift plain extends across northern Clinton County. Present mapping (Fig. 4) west of the Goose Lake Channel is similar to that of Udden (1905, p. 417) and summarized by Kay and Apfel (1928, opp. p. 15). The most significant revision is the recognition of a narrow strip of Kansan drift between the eastern limit of the Iowan drift and the Goose Lake Channel.

The present study shows that in an area of approximately 25 square miles east of the Goose Lake Channel the surface material is Shelbyville. In sec. 11, T. 82 N., R. 5 E., 2½ to 3 feet of Shelbyville drift overlies Kansan gumbotil.

The Kansan drift plain is thoroughly dissected, has a relief of more than 100 feet, and is covered with loess that averages 10–15 feet thick on the divides. The Kansan drift plain is more dissected, and the drift is more deeply leached than the Illinoian drift, and is much more dissected than the Iowan or Shelbyville drift plains. The latter, except near the border, are covered with thin loess or the loess is absent.

The oxidized and leached portions of the Kansan drift are chocolate brown to yellow; the unoxidized and unleached portions are blue gray, jointed, and compact.

An outstanding but not unusual feature of the Kansan drift in this area is the presence of a distinctive reddish-brown ferretto zone that is developed on the drift beneath loess, where gumbotil did not develop or was removed by erosion.

A noteworthy section of the Kansan drift which shows the development of silttil as defined by Leighton and MacClintock (1930, p. 41) is present in a road cut in SE¼ of SE¼ of sec. 12, T. 82 N., R. 6 E. The silttil lies above a 2–3 foot ferretto zone and is overlain by Farmdale loess which in turn is overlain by Peorian loess. In other places the gumbotil has been changed to second-cycle silttil.

Illinoian Drift in Iowa

Norton (1899, p. 492) indicated Illinoian drift overlain by loess over about three-fourths of Scott County. Exposures and auger borings indicate Illinoian till overlain by Farmdale loess, which is overlain by Peorian loess, on the south bluffs of the Wapsipinicon River in sec. 34, T. 80 N., R. 5 E., northwest of Princeton.

Udden (1905, p. 418) suggested that the Illinoian might exist in Clinton County, but found no proof of its presence. Carman (1909, p. 38) noted that a small area in southeastern Clinton County was probably covered by Illinoian ice.

Auger borings indicate two areas of Illinoian drift in Clinton County. One area lies between the southeastern margin of Iowan drift and the Goose Lake Channel (Fig. 4). Another area lies between the southwestern border of the Shelbyville drift and the Goose Lake Channel.

The Illinoian drift in Clinton County is more deeply weathered than the Iowan or the drift of the Green River Lobe. Depth of leaching ranges from 9 to more than 17 feet. The till has none of the ferretto zones so common in the Kansan.

Illinoian Drift in Illinois

Illinoian drift is present north of the Shelbyville from Fulton to Sterling (Fig. 4). Occasionally gumbotil has developed, and at many places the drift is overlain by Farmdale loess. The Illinoian occupies a higher, more dissected, and more heavily loess-covered area than the area of Shelbyville drift to the south.

Illinoian drift occurs at Cleveland, East Moline, Moline, Hampton, Rapid City, and as far north as sec. 17, T. 19 N., R. 2 E. Between Hampton and Rapid City gumbotil, overlain by Farmdale loess, with Peorian loess above, is present in the valley walls of the Mississippi River. Southwest of Hillsdale in sec. 36, T. 19 N., R. 2 E., compact till containing many fragments of coal is overlain by terrace deposits of sands, silts, and clays which have an altitude of 640 feet A.T.

Wisconsin Drift

Iowan

The Iowan drift of Clinton County, Iowa, was deposited by a long, narrow lobe of ice which extended across the central part of the county from the main body of ice to the west. The ice lobe, which extended across portions of southern Jones and northern Cedar counties before reaching Clinton County, was previously mapped eastward to the alluvial flat along the Mississippi River. Present mapping (Fig. 4) based on road-cut exposures and numerous auger borings indicates that the narrow lobe of Iowan ice did not extend east of the Goose Lake Channel. A narrow strip of Kansan drift lies between the Iowan drift and the west side of the Goose Lake Channel to the north while a small area of Illinoian drift lies between the Iowan drift and the western side of the Goose Lake Channel to the south.

Along its borders the Iowan drift lies beneath a thick cover of loess. Leaching of carbonates from the loess has generally proceeded to a depth of 7 or 8 feet, and calcareous loess overlies calcareous Iowan till. Away from the borders the loess is thin or absent and leaching ranges from 18 inches to 5 feet. A pebble band at the top of the Iowan is common throughout the area, and granite boulders are numerous and in some cases large. Iowan drift is present over bedrock, adjacent to Kansan, and adjacent to Illinoian, but at no place was the Shelbyville found above the Iowan.

Tazewell (Shelbyville) in Iowa

The Green River lobe of Shelbyville ice extended west of the Mississippi River to the eastern border of the Goose Lake Channel in Clinton County and northwestward to within about 1½ miles of the village of Goose Lake (Fig. 4). The ice crossed the Mississippi River between North Clinton (Lyons) and Princeton, Iowa, into an area previously mapped as Iowan and Kansan drift. The southwestern portion of the extension is separated from the Goose Lake Channel by a small area of Illinoian drift.

A critical exposure in the above interpretation occurs in sec. 11, T. 82 N., R. 5 E., where calcareous Shelbyville till overlies Kansan gumbotil (Fig. 4).

Tazewell (Shelbyville) in Illinois

North of Geneseo as far as Fulton, the surface material is Shelbyville drift, except where it is covered by later alluvium in the wide lowlands of the Rock River valley, the Meredosia Channel, and the Cattail Channel (Fig. 4).

Shelbyville drift is present north of Geneseo at numerous places in the Cordova quadrangle —namely, Pink Prairie, Sharon, Spring Hill,

Hillsdale, south of Albany, east of Fenton, on the east side of Cattail Creek, and at Union Grove. The northern boundary of the Shelbyville extends from Fulton across the southern portion of the Morrison and Sterling quadrangles where the eastern limit of this study joins the boundary determined by Templeton (1939, unpub. thesis, Univ. Ill., pl. II).

The Iowan and Shelbyville drifts are similar in several respects. Where the loess is very thin or absent, the depth of leaching of the Shelbyville till ranges from 2½ to 6 feet as it does in the Iowan till under similar conditions. Where the Shelbyville drift is covered with 10 or more feet of loess, leaching of carbonates from the loess has proceeded for 8 or 9 feet, and calcareous loess overlies calcareous till. Where Iowan drift is under similar loess cover, the depth of leaching in the overlying loess is similar.

The Iowan and Shelbyville drifts are dissimilar in several respects. No pebble band was found on top of the Shelbyville, and the numerous, large, granitic boulders characteristic of the Iowan were not found in the areas of Shelbyville drift.

Drainage History

General Features

In the area between Fulton and Keokuk, two drainage systems operated during both the Sangamon interglacial stage and the Iowan substage (Fig. 5).

The northern drainage flowed through the Meredosia Channel southeastward to the Illinois River in Bureau County. A tributary which headed in a rock divide just west of Andalusia flowed eastward and northward along the course of the present Rock River valley to join the old Mississippi (Fig. 5). A divide also existed near Cordova from which a small stream flowed north to the old Mississippi and a stream flowed southwestward past Rock Island, Illinois, to join the stream in the lower Rock River valley.

The southern drainage system included the Iowa and Skunk rivers which flowed southward through the Keokuk rock valley. The old Iowa included a southward-flowing tributary which headed in the rock divide just west of Andalusia and received the waters of Copperas Creek, Eliza Creek, and Edwards River from Illinois and numerous tributaries from the west in Iowa.

Glacial Lake Milan

With the onset of Tazewell (Shelbyville) glaciation, the first blocking of the northern drainage system occurred at the "big bend" of the present Illinois River (Fig. 5). Here the ice ponded the old Mississippi to form the first stage of a glacial lake, herein called Lake Milan (Fig. 6). The lake, which in the initial stage extended from the Illinois River to the outlet over the rock divide west of Andalusia, was later restricted to the lower Rock River valley when the ice reached the western walls of the present Rock River valley. The outlines of the first stage of Lake Milan are not shown because the advancing ice progressively covered the area between the "big bend" and Geneseo. The bedrock topography map (Fig. 2) gives the general outlines of the area inundated.

Glacial Lake Cordova

When the Tazewell (Shelbyville) ice reached the west wall of the Rock River valley southeast of Cordova, glacial Lake Cordova was formed (Fig. 6). Glacial Lake Milan was then restricted to the lower Rock River valley and received meltwater from the ice, which may have already reached Geneseo, Illinois, or was advancing toward Geneseo from the east. The diagram of Lake Cordova, based on present topography, represents the earliest minimum extent and lowest level (641 ft. A.T.) which occurred promptly after the ice blocked the old upper Mississippi flow into glacial Lake Milan. The level of glacial Lake Cordova subsequently rose about 15 feet, the waters spilled over the divide near Cordova and cut the Cordova gorge. The relationship between this cutting and the Green River lobe was pointed out by Leighton (1923, p. 281). The initial cutting was undoubtedly accomplished quickly because the material removed was Illinoian drift which is at least 50 feet thick in this area. Exposures in the valley wall near Hampton and Rapid City, Illinois, show that gumbotil developed on the upper surface of the Illinoian till at an elevation of about 655 feet above sea level. The bedrock

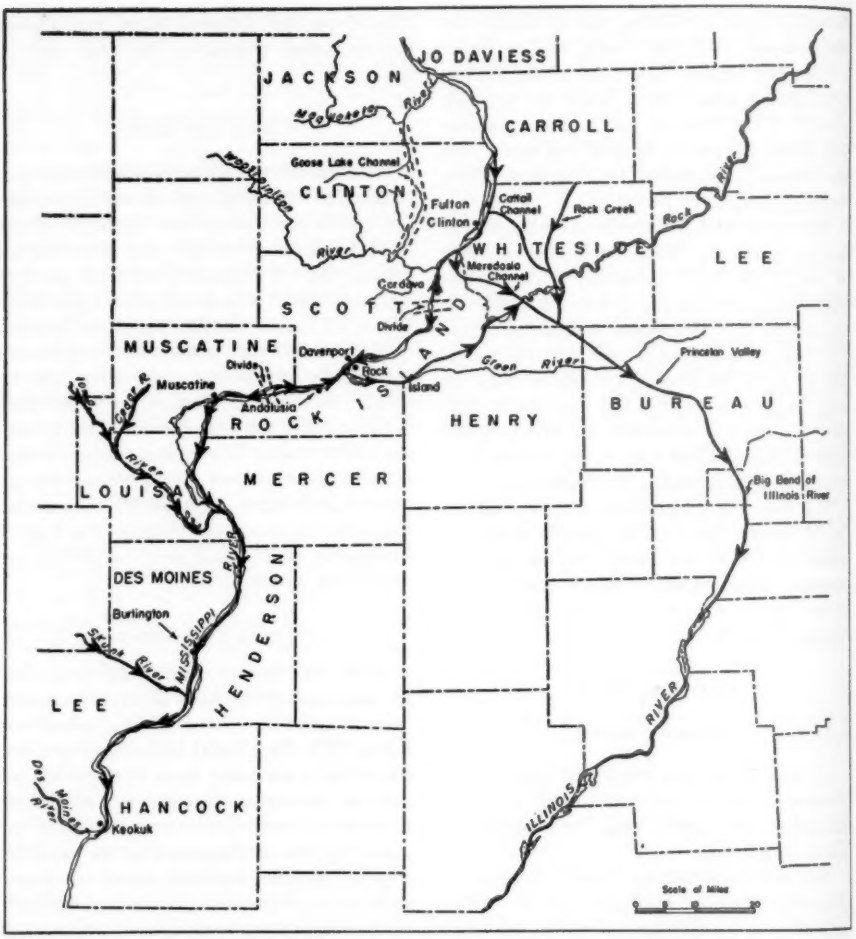


FIGURE 5.—DRAINAGE PATTERN DURING SANGAMON AND IOWAN

surface is at 600 feet or a few feet less in the valley. The highest terrace deposits of Lake Milan are found at 641 feet, thus the maximum height of the outlet of the lake was near 641 feet in elevation. About 15 feet of cutting through the Illinoian till therefore established southward flow through the Cordova gorge.

Glacial Lake Savanna

When the advancing Shelbyville ice crossed the Mississippi valley into eastern Iowa, glacial Lake Savanna (Leighton and Shaffer, 1949, p. 1904) formed north of Fulton. The outflow of this lake was by way of the Maquoketa River bay, through the Goose Lake Channel to the Wapsipinicon, and through the Cordova valley. The diagram for Lake Savanna (Fig. 7) is based on present topography and on the highest terrace deposits found (679 feet A.T.).

The records of this lake are found in sediments that rise to the approximate level of the low col in the Goose Lake diversion channel 10 miles west of Clinton, Iowa. The following section, discovered and described by Leighton, is exposed in a side hill pit in the city of Savanna,

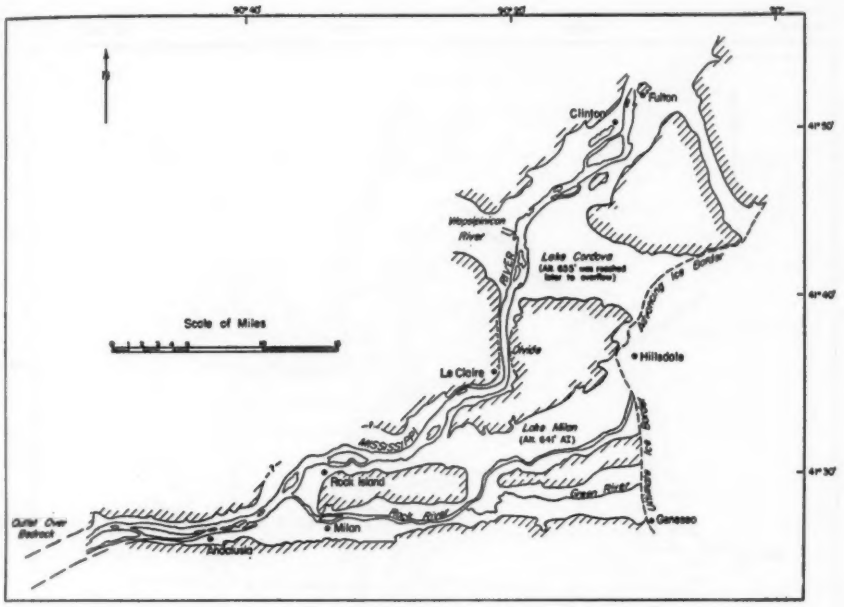


FIGURE 6.—GLACIAL LAKE MILAN (RESTRICTED) AND LOW-WATER STAGE OF GLACIAL LAKE CORDOVA
Generalized from present topography

northeast of the intersection of Gilbert and Pierce Streets.

Top of Section 699 feet A.T.	*Thickness (Feet)*
Loess, calcareous	20
Bedded deposits of variegated silts, sands, and clays. The sand is rusty streaked, the remainder blue, gray, and maroon. All calcareous	8
Green clay shale (Maquoketa)	2 (exposed)

Another side hill excavation occurs at the same elevation about 300 feet west and northwest of this section. In the southwest corner of this excavation there is more than 4 feet of bedded clay containing lime concretions, overlain by 1½ feet of fine sand and silt with clay partings. All the material is calcareous and is overlain by calcareous loess.

In the NE¼ of sec. 8, T. 23 N., R. 4 E., along the east-west road through the section, there is an excellent exposure of water-laid sediments. The exposure includes more than 20 feet of horizontally bedded sands, rusty near the top, with layers of silt 1–2 inches thick in the upper 2 feet. Limy streaks and lime concretions are prominent in the top 4 feet. At the base is a layer of rubble composed of fragments of the bedrock below. The top of the bedded deposits is between 675 and 680 feet above sea level.

Northeast of Fulton, on a small tributary to Otter Creek, in the NW¼ of sec. 17, T. 22 N., R. 4 E., the following materials are exposed:

Top of Section 650 feet —660 feet A.T.	*Thickness (Feet)*
Loess, sandy, fossiliferous in lower portion, leached to a depth of 2–3 feet	15
Sand, with fine sand and a few silt beds. Arranged like foreset delta beds	40

The water-laid sediments in the sections are too high for the "Mankato" terrace deposits in the Mississippi valley and its tributaries and, unlike the "Mankato" terrace deposits, are covered with rather thick loess.

The waters of Glacial Lake Savanna joined those of glacial Lake Milan (restricted), flowed through the rock valley west of Andalusia, and lowered the outlet fairly rapidly. With the

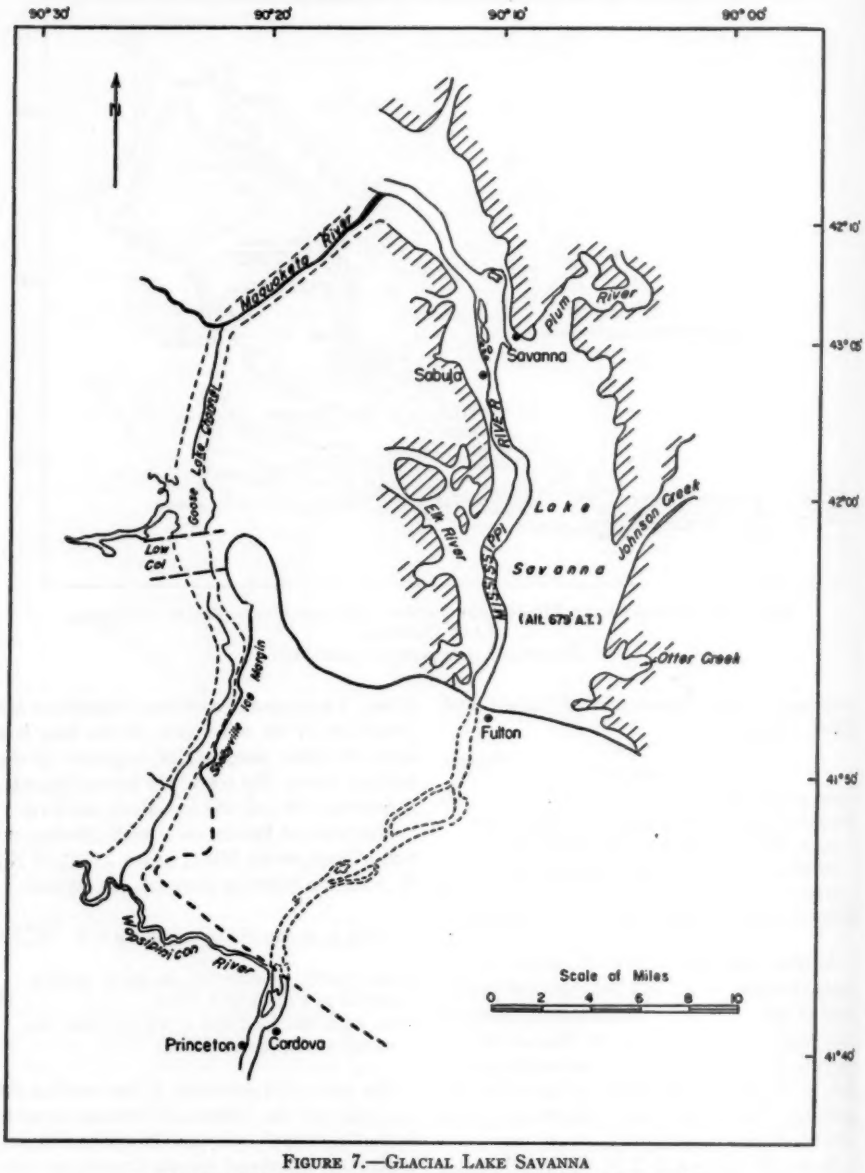


FIGURE 7.—GLACIAL LAKE SAVANNA
At first, highest-level stage. Generalized from present topography

lowering of this outlet, the intrenchment of the Cordova gorge and the withdrawal of the Green River lobe to the east, the two drainage systems were integrated to form the present Mississippi River valley between Fulton and Keokuk.

TAZEWELL TERRACES

Two sets of loess-covered Tazewell terrace remnants, deposited under different controlling conditions, are present between Geneseo, Illinois, and Keokuk (Fig. 8).

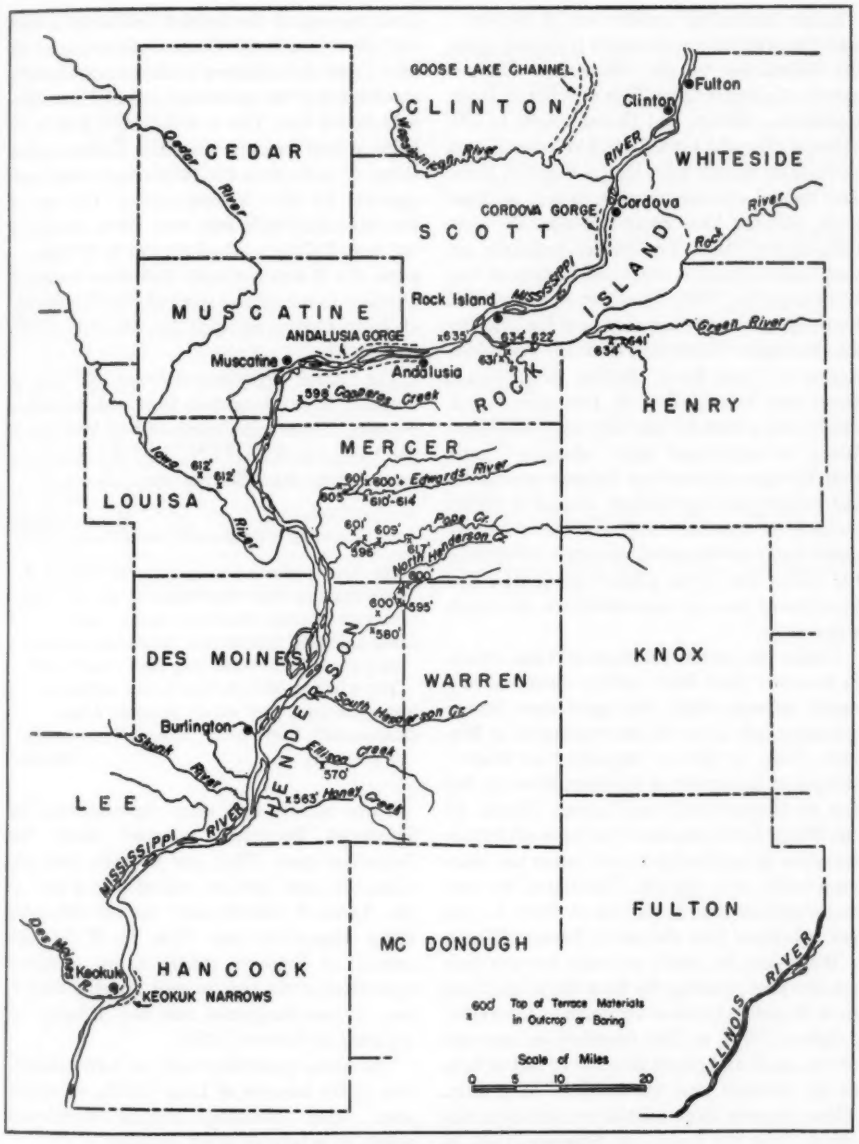
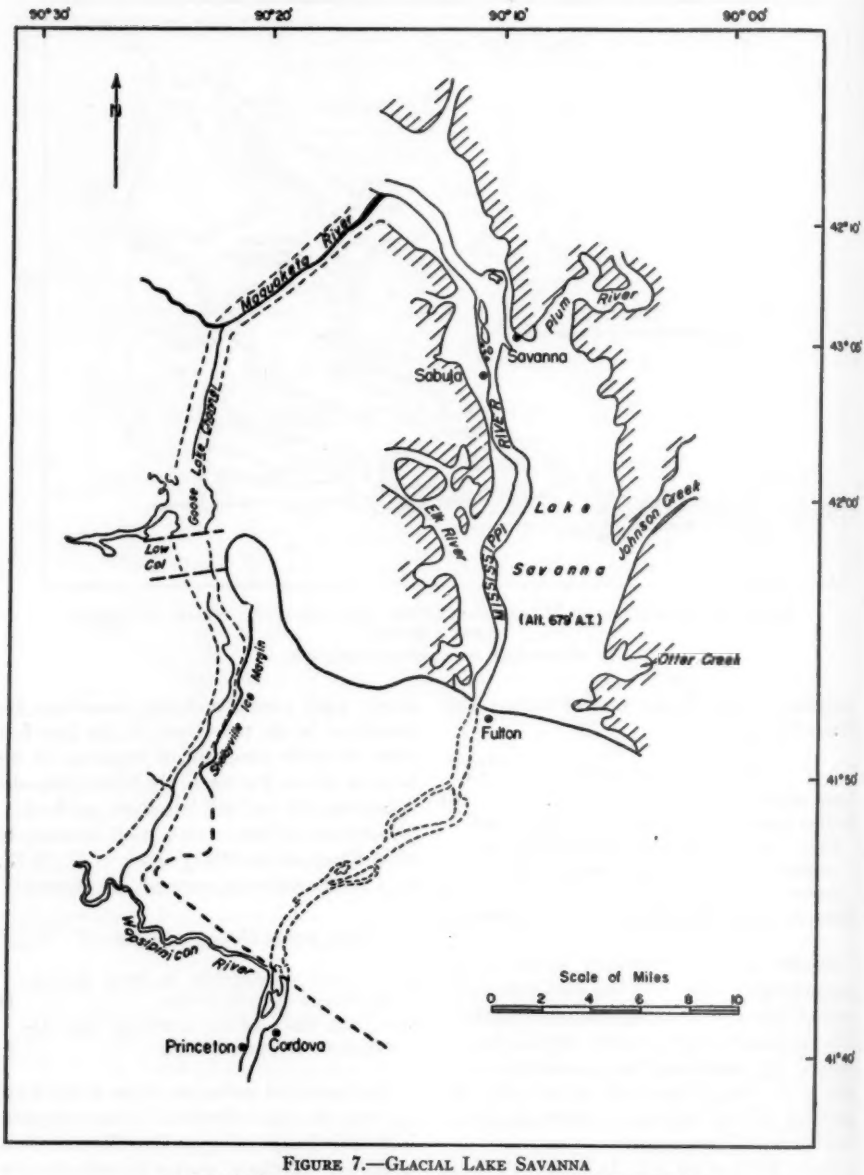


FIGURE 8.—TAZEWELL TERRACE REMNANTS

One set north of the Andalusia rock divide in the lower Rock River valley extends from Mineral Creek about 6 miles west of Geneseo, westward beyond the mouth of the Rock River to Linwood, Iowa, about 2 miles northeast of Andalusia. The materials of the terrace remnants are slack-water fine sands and silts covered with loess. The terrace remnants vary in height, but the higher, undegraded ones rise to an elevation of about 640 feet A.T. They resulted from slack-water deposition in glacial Lake Milan.



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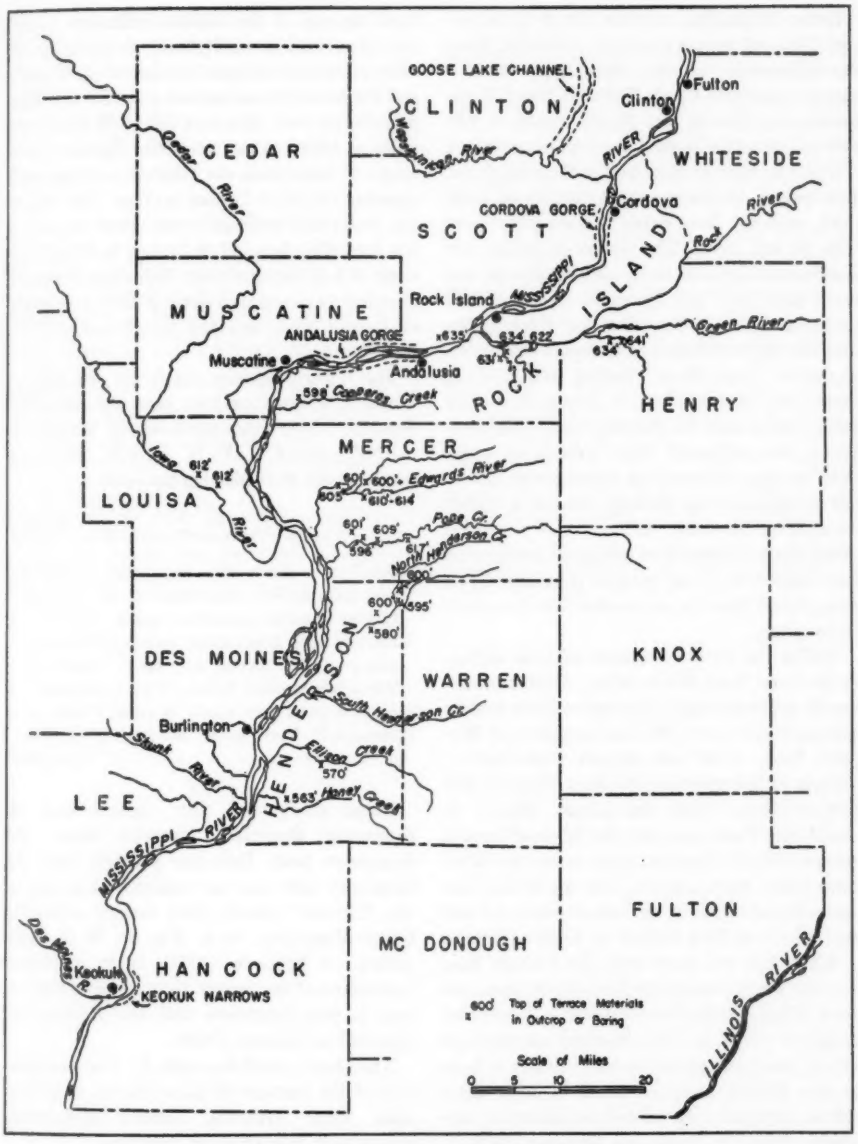

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Below Muscatine, another set of loess-covered Tazewell terrace remnants is present along the tributaries to the Mississippi River—namely, Copperas Creek, Edwards River, Pope, Henderson, Ellison, and Honey creeks in Illinois and along the Iowa River 3 miles northwest and also for several miles east of Wapello, Iowa. The bedded sediments are 8–10 feet or more thick, and the loess cover is commonly more than 10 feet thick. The terrace remnants are slack-water deposits composed largely of fine sands and silts. Their deposition was controlled by a constriction in the valley near Keokuk. Before the upper Mississippi drainage was added to that of "Iowa River" flowing past Keokuk, there was through-flow in the narrow rock valley, and a norm for that flow was established. When the additional large volume of water from the upper Mississippi drainage was added, the constriction at Keokuk caused a higher water level upstream with areas of relatively quiet water in expanded marginal portions of the valley and in the ponded tributaries. The fine-grained deposits accumulated in the quiet-water areas.

Unlike the terrace remnants of Lake Milan, in the lower Rock River valley, which rise to a nearly uniform height, the top of these terrace remnants falls from 612 feet northwest of Wapello, Iowa, to 601 feet opposite New Boston, Illinois, at the mouth of the Iowa River, to 563 feet on Honey Creek near Lomax, Illinois. At the Honey Creek exposure the Tazewell terrace merges with and begins to pass under the Mankato valley train deposits. The top of the bedded sediments falls at the rate of about 1.1 feet per mile from New Boston to Lomax, Illinois.

Water-laid fine sands and silts beneath loess are also present along the Iowa River upstream from Wapello at least as far north as Iowa City. Leighton (1916, p. 136) described an exposure of silt, sand, and gravel beneath 10 feet of loess at the Helman sand pit north of Iowa City. These deposits were related to meltwater discharge from the Iowan ice. Schoewe (1924, p. 147) described the high terrace of Lake Calvin (Fig. 3) at about 680 feet above sea level in the vicinity of Iowa City. He computed the slope of the top of the terrace at 1.4 feet per mile.

With a surface elevation of 680 feet at the Helman sand pit and a thickness of 10 feet of loess the top of the bedded sediments is 670 feet above sea level. Helman pit is about 40 miles from the exposure northwest of Wapello and the top of the water-laid material near Wapello is 612 feet. This is a fall of 58 feet in 40 miles or 1.45 feet per mile. The Helman pit is about 50 miles from the Mississippi valley wall opposite the New Boston section. The top of the water-laid sediments near New Boston is 601 feet. This is a fall of 69 feet in 50 miles, a slope of 1.38 feet per mile. This slope is almost identical to the surface slope of the high terrace of Lake Calvin reported by Schoewe (1924, p. 147).

The terrace deposits northwest and east of Wapello, lie on Sangamon peat and floodplain deposits. Three miles northwest of Wapello in the NE¼ sec. 5, T. 74 N., R. 3 W., the exposure contains the following materials.

Top of Section 620 feet A.T.	*Thickness (Feet)*
Loess, brown above and gray below. Lower 3 feet contains lime concretions at the top and tubular limonite concretions below. Leached	8
Sands and silts, interbedded yellow-brown sands and gray silts, become very sandy downward. Top 4 feet leached, bottom 4 feet calcareous	8
Sangamon peat, very woody in upper 2 feet	3
Gumbosand? Very sandy material gumbo-like in character	1 (exposed)

These sandy and silty deposits may be Sangamon floodplain material above the Sangamon peat. They are younger than the Sangamon peat and are regarded as a part of the Tazewell (Shelbyville) terrace remnants below Muscatine, Iowa (Fig. 8). If the high terrace of Schoewe (1924) is an upstream equivalent of the terrace near Wapello then it, too, is post-Sangamon and not Illinoian as regarded by Schowe (1924).

The above possibility calls for a re-examination of the terraces of Lake Calvin, especially since other evidence, already mentioned, points to a lake of short duration.

The high terrace of Lake Calvin may be an upstream continuation of the Tazewell terrace near Wapello, and the intermediate and lower terraces a continuation of the upper and lower "Mankato" terraces which are widely recognized in the Mississippi valley. If the high

terrace of Lake Calvin proves to be equivalent to the Tazewell (Shelbyville) then the Shelbyville portion of the Tazewell and the Iowan would be contemporaneous, and on the basis of priority Shelbyville would be redefined as Iowan. These matters merit further field investigation.

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POSTGLACIAL MARINE SUBMERGENCE IN CENTRAL ARCTIC CANADA

By J. Brian Bird

Abstract

In the Canadian Northwest Territories west of Hudson Bay, the postglacial marine transgression was more extensive than in any other part of North America. Although there are numerous isolated observations of strand lines and other raised marine features, there has been no systematic study of the maximum depth of this sea and the form of the subsequent emergence of the land.

Measurements made during 3 summer's field work in the area show that on the mainland the sea reached a depth of 360–400 feet. On the islands in the north of Hudson Bay the depth was 550–650 feet. The difference is explained by unequal waning of the ice sheet. In the western part of the Thelon and Dubawnt basins a large proglacial lake preceded the invasion of the sea. The lake fell by clearly defined stages. Subsequent emergence from the sea was continuous until the final 100 feet when there were two periods of temporary stability. There is widespread evidence that the land is still rising.

In the areas that were submerged, unconsolidated debris has been reworked by wave action into beaches, spits, bars, and other shoreline features. Along the shallower coasts these have completely changed the character of the landscape. The marine transgression did not last long enough to produce erosional features in consolidated rock.

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Introduction

Nearly the whole coast line of Canada has been submerged by the sea in postglacial times. Widespread marine transgression occurred when the earth's crust failed to regain its former elevation after the depression produced by the weight of the Pleistocene ice sheet. Nowhere in Canada is the physiographic effect of this submergence clearer or more important than in the central Arctic around the northern end of Hudson Bay. One of the first explorers to visit Arctic North America, Captain Luke Foxe, wrote (1635, p. 329): "I doe thinke that in Winter ... *there are* flowings of water incredible to be beleeved ...; for it may be perceived that the water hath been upon the land and islands higher by five or six fathoms than the usual spring tides." Although Foxe underestimated the height of submergence and when it occurred, it had impressed him greatly. Many subsequent explorers have recorded the overwhelming evidence of extensive marine submergence in postglacial times. Such easily recognizable evidence includes beaches, in all respects identical with modern beaches except that they are many miles from the present coast; marine shells 300 to 400 feet above sea level but otherwise similar to those species found on the modern foreshore; and whale skeletons many miles inland. Even the Eskimos are aware that the sea was once higher than at present.

The significance of the postglacial marine submergence in the evolution of the landscape of central Arctic Canada has been stressed by many geographers and geologists, and some have gone so far as to divide the area into physiographic regions on the basis of submergence in postglacial times (Tyrrell, 1898, p. 159; Robinson, 1944, p. 3). It is, therefore, surprising that, in comparison with Scandinavia, the analogous area in Europe, our knowledge of the maximum height of the submergence, its extent, and the form of the subsequent uplift is very meager.

The first scientific observations on the depth and extent of the postglacial sea were made by Parry (1824), while he was seeking a northwest passage by way of Hudson Bay. All the more important observations have, however, been made during the past 60 years. They began with the survey of central Keewatin by Tyrrell in 1893 and 1894, were continued by members of the 5th Thule expedition, notably T. Mathiassen (1931; 1933), and further extended by T. H. Manning (1942) and members of his parties between 1933 and 1938. The results have been summarized up to 1930 by Cooke (1930), and there are later additions by Washburn (1947, p. 61–73). The majority of these observations are incorporated on a map published by The Geological Society of America (Flint *et al.*, 1945). This map shows only six separate observations north and northwest of Hudson Bay. The present author has recently spent 3 field seasons in the central Canadian Arctic and has been able, while making geographical surveys for the Geographical Branch, Department of Mines and Technical Surveys, Ottawa, to collect additional data on the postglacial submergence. In 1948 a survey was made of Baker Lake and the Thelon River as far west as Beverly Lake. In 1950 the south and east coasts of Southampton Island were examined, and in 1952 the west coast of Southampton Island, Coats Island, and the mainland from Chesterfield Inlet to Repulse Bay, including Wager Bay. In addition, aerial photographs of the area were studied in detail.

Evidence for Submergence

Although evidence of past marine submergence is widespread close to modern sea level, near the upper marine limit the evidence is less apparent. As the maximum transgression of the sea was of short duration, erosional features at this height were cut only in unconsolidated material. These have since been destroyed rapidly by solifluction. Constructional marine landforms, including beaches, spits, and bars, are more common, if only because they formed more quickly in the early stages. They also become rarer near the upper marine limit. When only a fragment of a deposit survives, it may easily be confused with small kame terraces and other glacio-fluvial features. In addition, some forms of lateral moraines, particularly when seen from a distance, may be mistaken for strand lines. For these reasons high elevations claimed for the marine limit have not always been substantiated by subsequent investigation. River terraces, theoretically, should enable computation of the base level of their river, and hence in most cases the level of the sea when they were formed. In practice, terraces are rare (confined mainly to the east coast of Southampton Island), and nowhere are they found near the upper marine limit. The most serious disadvantage of using water-formed features to determine the upper limit of marine submergence is that distinction between levels formed by fresh-water lakes, brackish estuarine conditions, and the sea is difficult.

A second method of determining the marine limit is available using fossil marine shells. The presence of unbroken shells of marine organisms in superficial deposits is in almost all cases proof that the area has been invaded by the sea. The solitary shell found at unusual heights and positions may be suspect as having been carried by sea birds, but large numbers of unbroken shells in a deposit have been accepted as good evidence for submergence. They are found most frequently in "clay boils" (nonsorted stone circles) (Washburn, 1950, p. 11) formed by frost action from patches of marine clay deposited on the surface.

Shells were observed in quantity only where limestone is abundant. They are numerous on Southampton Island where they are found up to heights of 480 feet; on Coats Island where there is much Paleozoic limestone; and on the southwest side of Repulse Bay, where there is thick limestone-rich till. West of Hudson Bay,

where there is no limestone, fossil marine shells are rare and have not been found above 160 feet (Thelon Gorge). There is further uncertainty of the upper marine limit, as it is not known how soon after deglaciation marine animals invaded the area.

Difficulties inherent in these two methods have led the author to develop a third. When the ice sheet waned, drift from the ice was deposited on rock surfaces. In some areas this takes the form of a till deposit 100 feet or more thick; in others the deposit consists merely of isolated boulders, often perched on smaller pebbles and gravel. If the deposit is thin (6 feet or less) and rests on a steep slope, in many places it stops abruptly at a certain elevation, and the rock below is clear of material. Frequently a clear line separates perched boulders and ground moraine above, from bare rock below (Pl. 2, fig. 1). Stanley (1939) describes a similar disappearance of perched boulders, below a certain level, east of Hudson Bay. If the slope is less steep, the material below the line may not be removed entirely, but may show signs of sorting, whereas above it has a heterogeneous composition. This line is assumed to be the upper limit of water submergence in the area (Bird, 1953, p. 21, 24). Although the method has some of the drawbacks of the other two, it has two great advantages. The demarcation line between graded and ungraded deposits may be found easily in the Precambrian rock (but not limestone) areas of moderate relief, and, once found and measured accurately, the line is an absolute determination. With other methods there is always the possibility that higher beaches or shells may be found in a later investigation.

As might be expected, the three methods do not give identical results. On the east coast of Southampton Island where the relief is greatest, the highest marine shells are about 80 feet lower than the highest strand line, which in turn is about the same distance below the upper marine limit as inferred from perched boulders. This spread of about 160 feet compares favorably with a spread of 180 feet found by Nichols (1936, p. 6), using somewhat different methods, in an area of similar relief near Cape Wolstenholme on the south side of Hudson Strait. On the Keewatin mainland, where the relief is in general more subdued, the upper strand lines and the upper limit as inferred from the ground moraine normally coincide and are about 200 feet above the highest shells.

Depth of Submergence

When determinations of the maximum altitude of water submergence in the area are examined, it is immediately evident that there are very large blanks on the map, notably on both sides of Roes Welcome Sound where the coastal plain is so low that it was totally submerged, and in the basins of the Quoicht and Back rivers, which still await scientific examination. There are now sufficient data, however, to make valid inferences.

In the west of the area (upper and middle and Dubawnt basins) the submergence reached 740 feet and may have exceeded 800 feet. In the center of the area (lower Thelon Basin, Baker Lake, Wager Bay) it ranged from 360 to 400 feet; and in the east (Southampton and Coats Islands, Melville Peninsula) the submergence was from 550 to 650 feet. The most characteristic feature of these figures is the wide variation in height and particularly the sudden drop of nearly 400 feet between the first and second group. Fortunately the explanation is simple. The wave-cut cliffs, well-formed beaches at a few altitudes (Pl. 1, fig. 1), and overflow channels to the north joining the Thelon Basin to the upper Back River and from there to the Arctic Ocean all indicate that a glacial lake rather than the sea formed the higher submergence features of the west. The glacial lake was formed when the normally east-flowing drainage of the deglaciated western area was blocked by ice remaining in eastern Keewatin. The resultant lake probably exceeding 30,000 square miles in area was comparable in size to Lake Superior.

The rather similar difference in height between the mainland and the islands may be due to one of two causes. The plane of maximum submergence may have been subsequently warped, rising toward the east and southeast. The data from Wager Bay hardly supports this hypothesis to the required degree of tilt. The alternative is that the mainland area was occupied by an ice lobe separate in form and movement from the ice in the east and which survived after the islands were clear of ice.

The glacial geology supports this, but the evidence is not conclusive.

There are no field observations from the region around the lakes of the middle Back River. Aerial photographs show no evidence of raised beaches and other emergent features, which are so striking around the Thelon Lakes. The marine limit (of transgression from the Arctic Ocean) is found in this region 50 miles to the north on the Arctic Slope.

TABLE 1.—MAXIMUM HEIGHT OF POSTGLACIAL WATER SURFACES

Location	Height in feet	Method of Determination*	Observer†
Keewatin Mainland			
Repulse Bay (west)	Over 475	Ground moraine	
	430	Strand line	
(east)	560	Strand line	Mathiassen
Wager Bay	362–401	Numerous determinations using strand lines and ground moraine	
Baker Lake	Approx. 390	Strand line	Tyrrell
Schultz Lake	355–375	Strand line and ground moraine	
Aberdeen Lake (east)	560	Strand line	
(center)	Over 740	Strand line and ground moraine probably exceeded 800	
Beverly Lake	Over 640	Strand line	
Dubawnt Lake	740	Strand line	Tyrrell
Southampton Island‡			
Liver Creek	Over 590	Ground moraine	
	480	Marine shells	
Mathiassen Brook	590	Ground moraine	
Cape Comfort	625	Ground moraine	
	470	Marine shells	
Coats Island			
Northeast end	Over 655	Ground moraine	

* The term ground moraine implies that above the height determination ground deposits are unsorted and contain undisturbed perched boulders, whereas below the height the material is water-sorted.

† Where no observer is listed, the determination was made by the author.

‡ For lack of supporting evidence this table does not include a determination of nearly 1200 feet made by Mathiassen (Freuchen and Mathiassen, 1925, p. 560) on Southampton Island.

RETREAT OF POSTGLACIAL SEA

Not only is the maximum extent and height of the submergence significant in the development of the landscape, but also the manner in which the water level fell, because during this period the land between the upper water limit and modern sea level was subject to wave action. In the west, the proglacial lake fell by well-defined stages, generally dropping by about 100 feet at a time, until, when it reached an altitude that is about 360 feet above present sea level, the sea invaded the area. Suddenly at this height the well-defined beaches are replaced by small, poorly developed storm ridges and fragments of true beaches only 1 to 6 feet apart vertically. In the other areas, where there was only marine submergence, similar forms are observed—storm ridges, bars, and

PLATE 1.—LAKE AND MARINE EMERGED STRAND LINES VIEWED FROM THE AIR

FIGURE 1.—STRAND LINES ON NORTH SIDE OF ABERDEEN LAKE

Above 360 feet they are few in number, well developed, and lake-formed. Below 360 feet they are marine beaches (Photograph courtesy of RCAF).

FIGURE 2.—OFFSHORE BARS FORMED IN POSTGLACIAL MARINE TRANSGRESSION

Now raised above sea level. South coast Southampton Island. Photograph is approximately 4 miles wide (Photograph courtesy of RCAF).

FIGURE 1

FIGURE 2

LAKE AND MARINE EMERGED STRAND LINES VIEWED FROM THE AIR

FIGURE 1

FIGURE 2

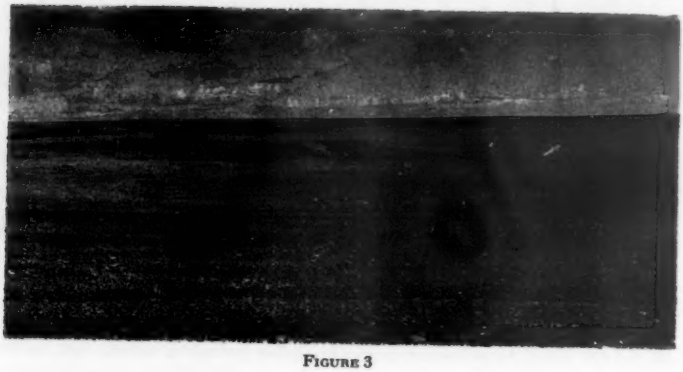

FIGURE 3

EMERGED STRAND LINES VIEWED ON THE GROUND

beaches formed and preserved almost fortuitously during the retreat of the sea; nowhere can evidence of a stationary water plane be observed. During the retreat of the sea there were no periods of stillstand at the upper and middle levels. The retreat was uninterrupted, though possibly of varying rate.

At the lowest levels (within 100 feet of the sea today) individual raised beaches are better developed and indicate at least two periods of temporary stability in water-land relationship in the most recent period of uplift. Such definite strand lines may be traced in Wager and Repulse bays, Roes Welcome coasts, Southampton Island, and Coats Island, where they are separated by a 20–30 foot vertical interval. The lower is about 35 feet above sea level. This coincides with the period of maximum sea temperature in the area as inferred from fossil shells (Laursen, 1946, p. 52).

One of the most important achievements in Scandinavia and the Great Lakes region of North America has been to show the tilt of individual strand lines and hence the warping of the earth's crust. Nowhere in the Canadian Arctic has this yet proved possible. In the region under discussion the only datum plane recognizable over any considerable distance is the upper limit of submergence. As it is difficult, if not impossible, to show that this was formed contemporaneously within even short distances, tilt cannot be deduced from it. At the lower levels and in the west of the area where the lake beaches are found, difficulty in correlating beaches over many miles when they are absent from intermediate areas has invalidated all measurements, although in some of the determination tilts greater than 1½ feet per mile should have been detected if they exist.

Physiographic Effect

It is now possible to show with some accuracy the limits of the marine submergence in the whole area (Fig. 1). Subsequent to the retreat of the ice, nearly half the area was covered for some period by the sea, lakes, or in some cases by both. At the maximum extension of Hudson Bay it was connected to the Arctic Ocean by a strait at least 250 feet deep over Rae Isthmus and possibly by another strait leading from the head of Wager Bay to Chantrey Inlet. The physiographic effect of the submergence varies in different areas depending on the type of rock exposed. On the Mainland the erosive effect has been slight because of the resistant granitic Shield rocks. In the Thelon Basin, where there are sandstones and conglomerates, the stillstand of the glacial lake was sufficient for some bluffs to be cut, but the till was generally thick enough to protect the underlying rock. In the limestone areas of the islands, on the other hand, the sea seems to have rejuvenated old cliffs and scarps. In most cases, however, they have been buried by talus resulting from subsequent violent freeze-thaw action. The important modification to the landscape produced by submergence has been the reworking by wave action of glacial deposits and material deposited by the sea. In most places where there is a sloping surface the material has been washed down the slope from the highest level of submergence, a movement aided subsequently by solifluction. At lower levels much of the material has been thrown up into storm ridges. At their greatest development the ridges blanket slopes from as high as 500 feet down to modern sea level. They are typically formed of boulders, cobbles, and pebbles. The ridges are only 2–3 feet high, are separated by shallow hollows, and many are extremely difficult to cross on foot.

Any superficial deposits of finer texture, such as sand, are redeposited in beaches, spaced at irregular intervals in the vertical plane, the exact position being controlled by local factors. On gently shelving coasts, more particularly in the limestone regions, offshore bars developed in the postglacial sea. Before these bars could retreat to the shore line, as is their normal

PLATE 2.—EMERGED STRAND LINES VIEWED ON THE GROUND

FIGURE 1.—UPPER LIMIT OF POSTGLACIAL MARINE SUBMERGENCE IN WAGER BAY AT 385 FEET
It is marked by lower limit of ground moraine and small delta (at right).

FIGURE 2.—HIGHEST STRAND LINE AT 390 FEET
On south side of Wager Bay; 85 yards wide

FIGURE 3.—LOW STORM RIDGES ON FEATURELESS LIMESTONE TERRAIN
West side Southampton Island (Photograph courtesy of Geographical Branch, Ottawa)

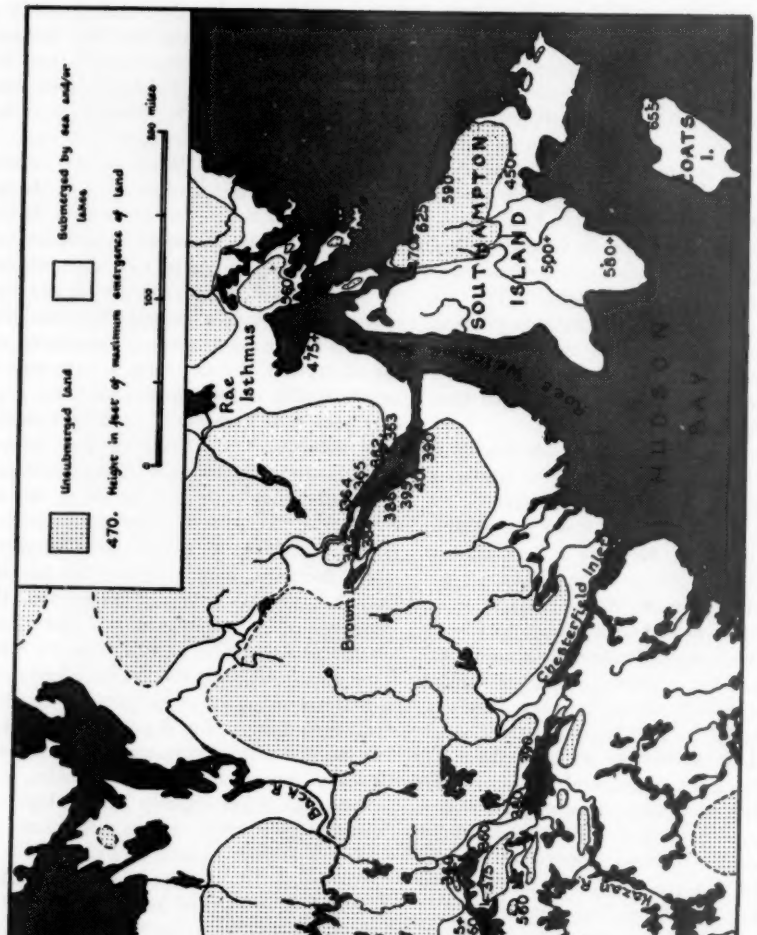

FIGURE 1.—EXTENT OF LAKE AND MARINE SUBMERGENCE IN CENTRAL ARCTIC CANADA

evolution, the sea fell, leaving them stranded. Subsequently another offshore bar would form, in turn to be stranded and so continuously to the present day, when the process is still in progress. Between one bar and the next a shallow lake may mark a former lagoon. The total result on the landscape is considerable modification (Pl. 1, fig. 2). It is, however, temporary and is already being destroyed by solifluction and other geomorphic processes.

Present-Day Upwarping

It is clearly recognized around the Gulf of Bothnia, in the region where the Scandinavian Ice Sheet was thickest, that land is still rising, at a rate of about 3 feet per century. It might be expected that the Hudson Bay region in North America, where the Laurentide Ice Sheet was probably thickest, is also rising. The evidence is conflicting and tends to be qualitative rather than quantitative. At the head of Baker Lake (which is connected directly with the sea) both the natives and white people say that the lake has retreated since the settlement was established in the 1920's. Farther south between Chesterfield Inlet and Eskimo Point a number of the older Eskimos agree that there are now rock islands where in their youth there were only reefs. From the distribution of archaeological sites there is no doubt that the land has risen at least 80 feet since the arrival of Eskimos with a Dorset culture, and 30 feet since the Thule peoples appeared. No dates are at present available for these cultures, but, if we assume that the Thule people arrived during the last 1000 years, this would give a rise since then of at least 3 feet per century. From the south of Hudson Bay evidence from the eighteenth century at Sloops Cove in Churchill Harbour has been interpreted as indicating both stability and a fall in sea level (Bell, 1896; Johnston, 1936). From James Bay it has been shown that the highest driftwood stands (of uncertain age) can be explained only by a fall in sea level of at least 10 feet (Manning, 1951).

The only quantitative data is from Churchill, where there has been a tide gauge since 1928. Gutenberg (1941, p. 747) analyzed the data for the first 10 years of operation and showed that for this period the land was rising at 3 to 6 feet per century. Since then the Canadian Hydrographic Office has thrown doubt on the validity of the earliest records, and an inspection of the data obtained since 1940 fails to show any positive trend. We are forced, therefore, to the conclusion that although continuing updoming, particularly around the north of Hudson Bay, is possible, it has not yet been proved. In this survey of the marine submergence of central Arctic Canada no mention has been made of the changes in sea level prior to the Wisconsin glaciation. The presence of valleys on the floor of Hudson Bay that have the form of normal subaerial river valleys implies a period of lower sea level than at present, while raised preglacial cliffs, marine benches, and river-cut rock terraces indicate periods of higher sea level. These fluctuations will also have to be unravelled before a full understanding of the evolution of the landscape in northern Canada is achieved.

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FIGURE 1.—EXTENT OF LAKE AND MARINE SUBMERGENCE IN CENTRAL ARCTIC CANADA

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Short Notes

AGE DETERMINATIONS OF THE SAN GABRIEL ANORTHOSITE MASSIF, CALIFORNIA

GEORGE J. NEUERBURG AND DAVID GOTTFRIED

Lead alpha activity measurements on zircon from the mafic border facies of the anorthosite massif in the San Gabriel Mountains, California, and from a granite pegmatite intrusive into it have yielded ages of 930 ± 90 million years and 810 ± 80 million years, respectively. The ages were determined in the U. S. Geological Survey laboratories using the method described by Larsen *et al.* (1952).

The zircon from the border facies of the anorthosite massif comes from the outer part of an ilmenite deposit in a much-altered norite in Sand Canyon, San Gabriel Mountains, Los Angeles County, California. The alteration is thought to be deuteric. The rock is a granoblastic aggregate of oligoclase, microcline, hornblende, biotite, ilmenite, pyrite, zircon, apatite, and carbonate in highly variable proportions. The anorthosite massif has been described by Higgs (in press). The zircon-bearing granite pegmatite is intrusive into the border facies of the anorthosite massif in Pacoima Canyon, roughly 10 miles east of Sand Canyon. The pegmatite may be related to granodiorite intrusives in the anorthosite in its eastern exposures. This pegmatite is briefly described by Neuerburg (in press).

The measurements on which these ages are based follow:

	α/mg/hr	Pb (ppm)	Age (million years)
Zircon—altered norite	18	7	930 ± 90
Zircon—pegmatite	50	17	810 ± 80

The alpha activity of the zircon from the altered norite is the lowest yet obtained for zircons separated from igneous rocks. This accords with recent investigations on the radioactivity of zircon from the rocks of the Southern California batholith. The zircon from the more mafic rocks, such as quartz diorite, is lower in alpha activity than zircon separated from granite. The alpha activity of zircon from pegmatite bodies follows no consistent trend.

Because of the low alpha activities and the slight lead content of the two zircons reported here, small errors in measurements would produce appreciable changes in the calculated ages. The data, however, are believed to be sufficiently accurate so that the ages are correct within 10 per cent of the reported figures. These are minimum ages for the anorthosite and pegmatite and demonstrate a Precambrian age for yet another anorthosite massif. Furthermore, the data provide a first age determination on the pre-Cretaceous crystalline rocks of the San Gabriel Mountains and point to the existence of Precambrian rocks near the edge of the continental shelf.

This work is part of a program conducted by the U. S. Geological Survey on behalf of the Division of Research of the U. S. Atomic Energy Commission.

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